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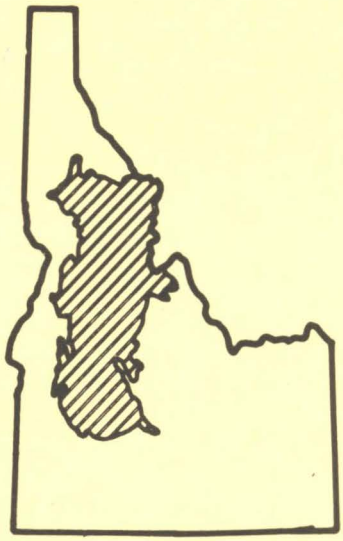
THE GENERAL GEOLOGY OF THAT PART OF THE NORTHERN ROCKY MOUNTAIN PROVINCE CONTAINING THE IDAHO BATHOLITH



by

EARL H. BENNETT II

for

U. S. FOREST SERVICE

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ABOUT THE AUTHOR

Earl H. Bennett II was born in Schenectady, New York, on September 19, 1942. After graduation from high school in New York, he served in the United States Air Force and spent a three-year tour of duty in Germany. Discharged in 1964, he enrolled at the University of Delaware with a major in Geology. After graduation in 1968, he continued his education at North Carolina State University at Raleigh, North Carolina, and received his Masters Degree there in 1970. Research at N. C. State was on recent beach sediments on the Carolina coast. In 1971, he enrolled at the University of Idaho in a doctoral program and received his Ph.D. in geology in 1973. Research at Idaho was conducted in the Idaho batholith. At the present time, he is employed as a research associate in the Idaho Bureau of Mines and Geology at Moscow, Idaho.

THE GENERAL GEOLOGY OF THAT PART OF THE NORTHERN ROCKY MOUNTAIN PROVINCE CONTAINING THE IDAHO BATHOLITH

THIS REPORT IS PART OF THE CHARACTERIZATION EFFORT FOR THE COOPERATIVE ADMINISTRATION - RESEARCH IDAHO BATHOLITH PROGRAM AND IS TO BE CONSIDERED AN APPENDIX TO THE BATHOLITH SOURCE BOOK.

by

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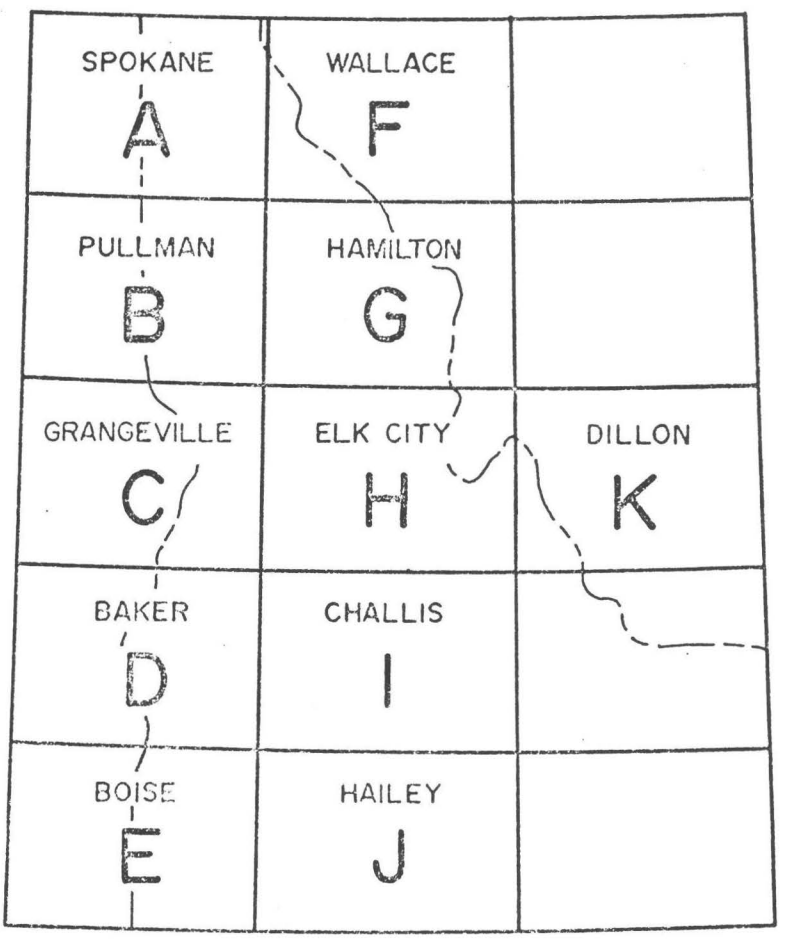
AMS SHEET - PLATE LEGEND



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INTRODUCTION

The area of the Northern Rocky Mountain Province that contains the Idaho batholith is shown in Figure 1. This province has been subdivided by the U.S. Forest Service into six sections (Figure 2) which are further divided into 79 subsections. The subsections are identified on overlays for the geologic maps which accompany this report.

PURPOSE OF REPORT

The purpose of this report is threefold:

1. Construct as detailed a geologic map as is practical at a scale of 1/250,000 for that part of the Northern Rocky Mountain Province which contains the Idaho batholith.

2. Describe the geology of the province at:

   a. The Province level

   b. The Section level

   c. The Subsection level

3. Discuss the lithology, weathering and jointing characteristics, drainage, topography, and geologic structure at the section and subsection level.

General geologic map of Idaho

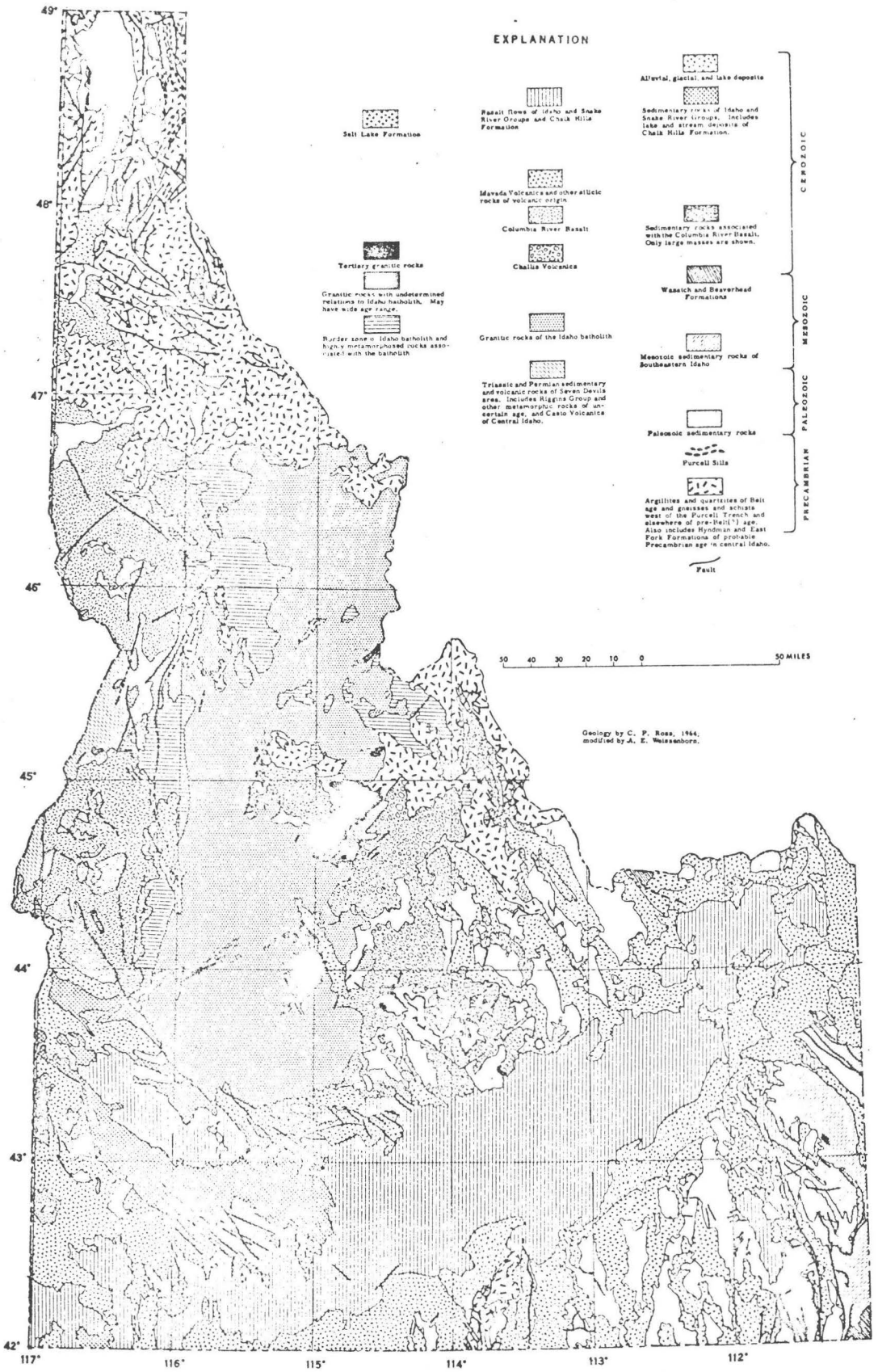

EXPLANATION
Alluvial, glacial, and lake deposits
Sedimentary rocks of Idaho and Snake River Groups, includes lake and stream deposits of Chalk Hills Formation.
Sedimentary rocks associated with the Columbia River Basalt. Only large masses are shown.
Wasatch and Beaverhead Formations
Mesozoic sedimentary rocks of Southeastern Idaho
Paleozoic sedimentary rocks
Purcell Sills
Argillites and quartzites of Belt age and gneisses and schists west of the Purcell Trench and elsewhere of pre-Belt(?) age. Also includes Hyndman and East Fork Formations of probable Precambrian age in central Idaho.
Fault
Basalt flows of Idaho and Snake River Groups and Chalk Hills Formation
Idavada Volcanics and other silicic rocks of volcanic origin
Columbia River Basalt
Challis Volcanics
Granitic rocks of the Idaho batholith
Triassic and Permian sedimentary and volcanic rocks of Seven Devils area. Includes Riggins Group and other metamorphic rocks of uncertain age, and Casto Volcanics of Central Idaho.
Salt Lake Formation
Tertiary granitic rocks
Granitic rocks with undetermined relations to Idaho batholith. May have wide age range.
Border zone of Idaho batholith and highly metamorphosed rocks associated with the batholith
CENOZOIC
MESOZOIC
PALEOZOIC
PRECAMBRIAN
50 40 30 20 10 0 50 MILES
Geology by C. P. Ross, 1944, modified by A. E. Weissenborn.
49°
48°
47°
46°
45°
44°
43°
42°
117°
116°
115°
114°
113°
112°

Figure 1

U.S. Forest Service Section boundaries (7) in the Northern Rocky Mountain Province

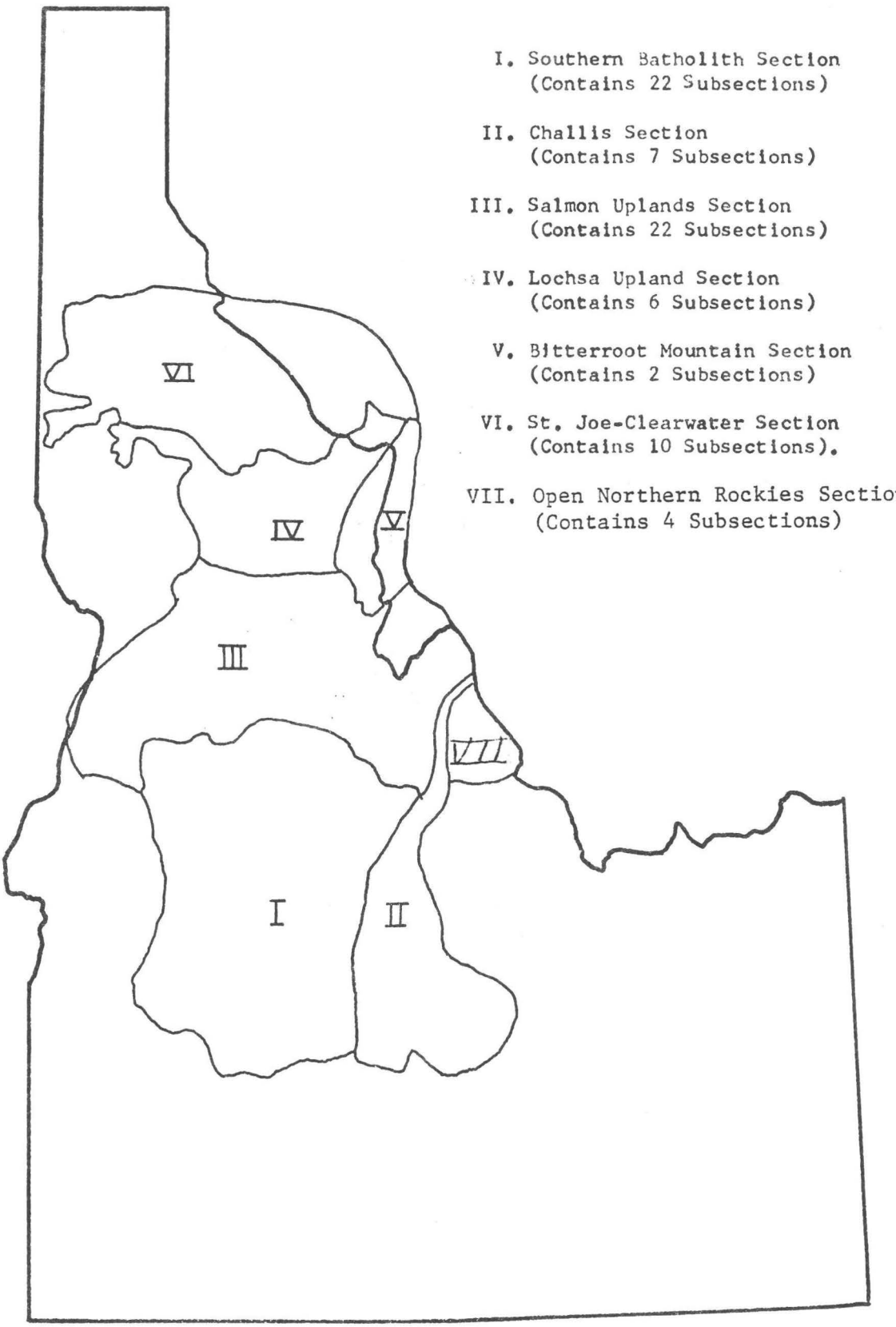


Figure 2

THE GEOLOGIC MAP OF THE PART OF THE NORTHERN ROCKY MOUNTAINS THAT CONTAINS THE IDAHO BATHOLITH

The geologic folio which accompanies this report consists of 11 Army Map Service (AMS), 1/250,000 scale maps that roughly cover the region from 43° to 48° north latitude and from 112° to 118° west longitude. The geology is limited to that within the state of Idaho (with a slight overlap into Montana). Accompanying the maps are 11 overlays which show the U.S. Forest Service section and subsection boundaries. Eleven black line copies of the AMS sheets are included which show the bibliographic data from which the geologic maps were compiled. Another set of 11 black line copies shows the distribution of linears which was compiled by Norman Day (University of Idaho) from ERTS (Earth Resources Technology Satellite) imagery.

The geologic maps are a compilation of data from more than 35 references. The geology on the AMS sheets is in a simplified form because of changes required to accommodate scale reduction from larger scale, more detailed maps. For example, major faults are shown on all the AMS sheets but on many maps, such as the Spokane sheet, faults were so numerous that lithologic contacts were obscured. In these cases many of the faults are just penciled in for clarity. For this reason the linear map and the geologic map should be examined together as most major faults show up in better detail on the linear map. Folds are discussed in the text but only major regional ones are shown on the maps.

The folio contains a legend showing all lithologic units, lithologic map unit symbols, and geologic structure symbols. Because there are a large number of lithologic units, the unit symbols are given in the text along with the unit name to facilitate map utilization.

THE GEOMORPHOLOGY OF THE NORTHERN ROCKY MOUNTAIN PROVINCE

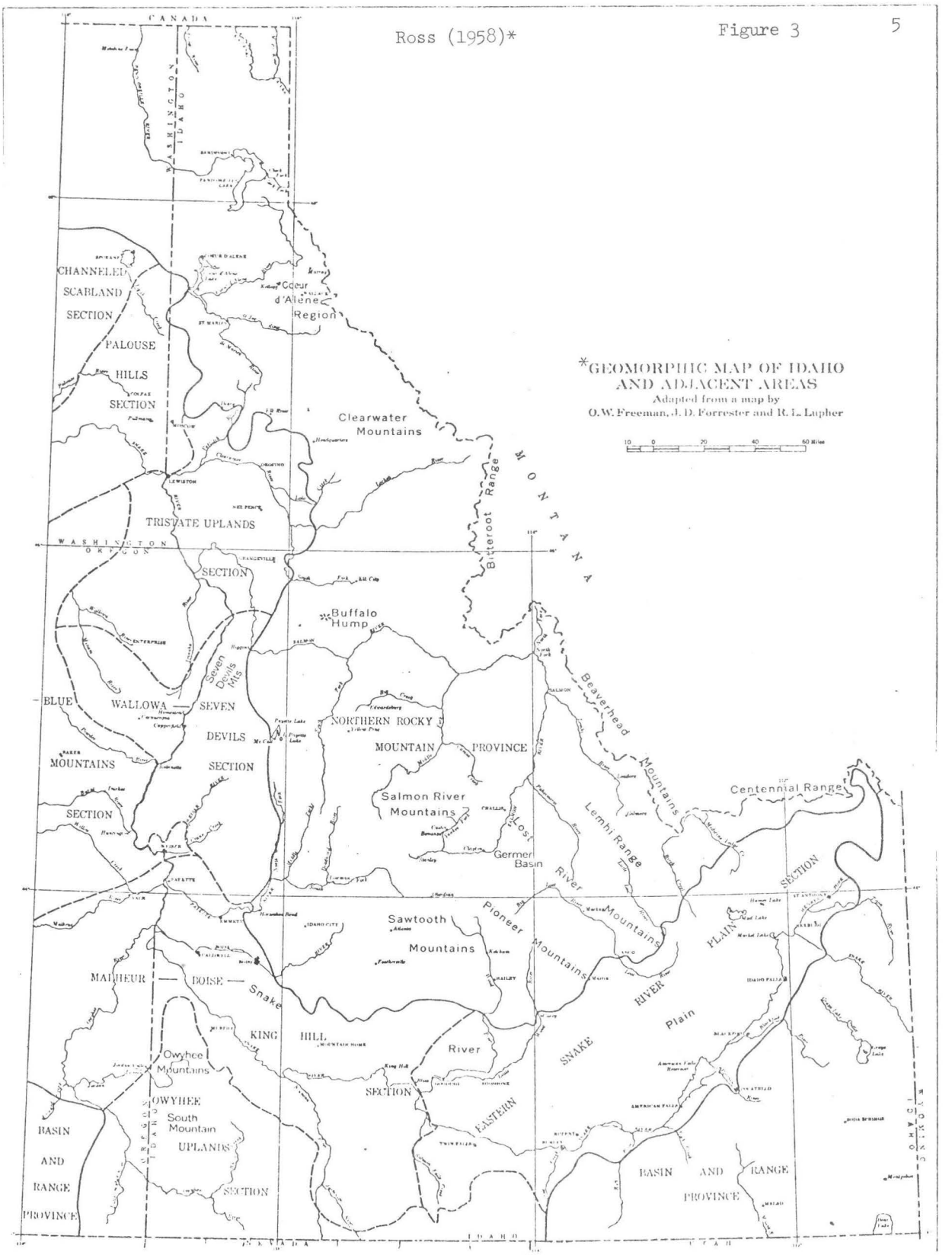
Ross (1958) summarized the geomorphology of the Northern Rocky Mountain Province. Figure 3 shows geomorphic boundaries. Most of the area contained within the province is underlain by Idaho batholith. The batholith is bordered on the west by the Tri-State uplands section and the Snake River section and on the south by the East Snake River Plain section and the Malheur-Boise-King Hill section.

Major drainageways in the province include the Snake River (southern portion), the Salmon River (central portion), and the Clearwater River (northern portion) with their tributaries. The extreme northern portion of the mapped area is drained by the St. Joe and Coeur d'Alene River systems which, unlike the other major drainages which flow to the Pacific, feed Coeur d'Alene Lake.

There are six major mountain ranges in the mapped area; from south to north these are: the Sawtooth Mountains, the Pioneer Mountains, a small portion of the Lost River Range, the Salmon River Mountains, the Clearwater Mountains and the Bitterroot Range.

Ross (1958)*

Figure 3



*GEOMORPHIC MAP OF IDAHO AND ADJACENT AREAS

Adapted from a map by O.W. Freeman, J.D. Forrester and R.L. Lupher

10 0 20 40 60 Miles

THE GENERAL GEOLOGY OF THE NORTHERN ROCKY MOUNTAIN PROVINCE

A discussion of the geology of the mapped area at the province level can either be so complex as to be meaningless to later discussions at the section and subsection level or too simplified for a clear understanding of the general geology. For this reason the general geologic units within the province will be briefly examined and one of these units, the Idaho batholith will be described in some detail as the batholith accounts for about two thirds of the entire province. The other geologic units will be discussed in greater detail at the U.S. Forest Service section and subsection levels. Here a thorough discussion of the geology is more meaningfull because the section or subsection boundaries often coincide with geologic unit boundaries.

Figure 1 shows the distribution of the 10 gross geologic units within the mapped area. These 10 units are conveniently discussed under the heading of Sedimentary and Metamorphic rocks, Extrusive Igneous rocks, and Intrusive Igneous rocks.

Sedimentary and Metamorphic Rocks

There are four major units of sedimentary (and their metamorphic equivalent) rocks which are, on a gross scale, geographically separated from one another. These four units, for the most part, border the Idaho batholith. Pre-Cambrian sediments and their metamorphic equivalents belong to the Belt Supergroup and are localized in the northern section of the map area; Paleozoic sediments, ranging in age from Cambriam to Permian, are found bordering the batholith in the southeastern section of the map area; Mesozoic age sediments are widespread in the central western and southwestern sections of the province and are associated with the Seven Devils volcanics (Permo-Triassic). Cenozoic sediments are found in the southwestern section of the province and are mostly lake bed or alluvial fan deposits. Cenozoic sediments are also found interbedded with the various volcanic rocks (Columbia River basalts or Challis volcanics).

Extrusive Igneous Rocks

Four major groups of extrusive igneous rocks occur in the mapped area. These are the Upper Cenozoic age Snake River basalts and Columbia river basalts, the Lower-Middle Cenozoic Challis volcanics, and the Mesozoic (Permo-Triassic) Seven Devils volcanics. The Seven Devils volcanics typically are metamorphosed to greenstones.

Intrusive Igneous Rocks

There are three major groups of intrusive igneous rocks in the map area the most extensive of which is the Idaho batholith complex. Smaller Tertiary plutons are associated with the third group, vast complexes of Tertiary dike swarms.

CURRENT IDEAS ON THE ORIGIN AND EMPLACEMENT OF THE IDAHO BATHOLITH

The Idaho batholith is an igneous complex which outcrops, more or less continuously across Idaho, for about 250 miles in a north-south direction and for about 90 miles in an east-west direction (see Figure 1). It is a member of a chain of batholiths which extends along the western border of North America. The chain includes the Southern California batholith and the Sierra Nevada batholith in California, the Central Nevada intrusive belt, the Idaho batholith, the Colville batholith in Washington, and the Coast Range batholith complex of Canada and Alaska. Smith et. al. (1971) believes that the Sierra Nevada and the Idaho batholith are connected via the Central Nevada intrusive belt.

Larson et. al. (1958) and R. Armstrong (Per. Comm.) have dated the Idaho batholith at 90-110 million years which is Late Cretaceous. Discrepancies in age dates of the batholith probably reflect a general reheating in Miocene time which affected K/Ar values in biotites which are most commonly used for age dating (McDowell and Kulp, 1969).

The formation of the West Coast batholith complex is believed to be directly linked to Plate Tectonics. Figure 4 (Hietanen, 1973) illustrates how magma is generated where oceanic crust is subducted below continental crust. Figure 5 (Hyndman, 1972) is a more generalized diagram which shows the zone of oceanic crust formation (Pacific Rise), underthrusting of the continent by the basaltic, oceanic plate, and the subsequent generation of various igneous rock types.

Figure 6 (Moore, 1959) shows the "Quartz Diorite Line" which is located at the western edge of the West Coast batholith complex. The rocks generally become more felsic (granodiorite grading to quartz monzonite) eastward from the Quartz Diorite Line in all of the western batholiths. Moore, Grantz and Blake, (1963), showed that potassium and silica in the granitic rocks increased from west to east and calcium and magnesium decreased which agrees with the compositional trend. In addition to composition and chemistry Mcdowell and Kulp (1969) showed that the age of the batholith rocks varied systematically (older in the west and younger to the east).

Strontium isotope studies are based on the idea that rubidium 87, which decays to strontium 87, is more common in granitic rocks than in basaltic rocks. This is because rubidium 87 is about the same ionic size as potassium and therefore is found concentrated in more potassic rocks (i.e. granitic).

There are four basic ways in which granitic magmas are believed to form:

1. Differentiation of mantle (peridote composition) or oceanic crustal material which is derived from the mantle (basaltic composition).

Sketches suggesting structural and magmatic evolution of the postulated subduction zone in the northern Sierra Nevada (Hietanen, 1973)

ANNA HIETANEN

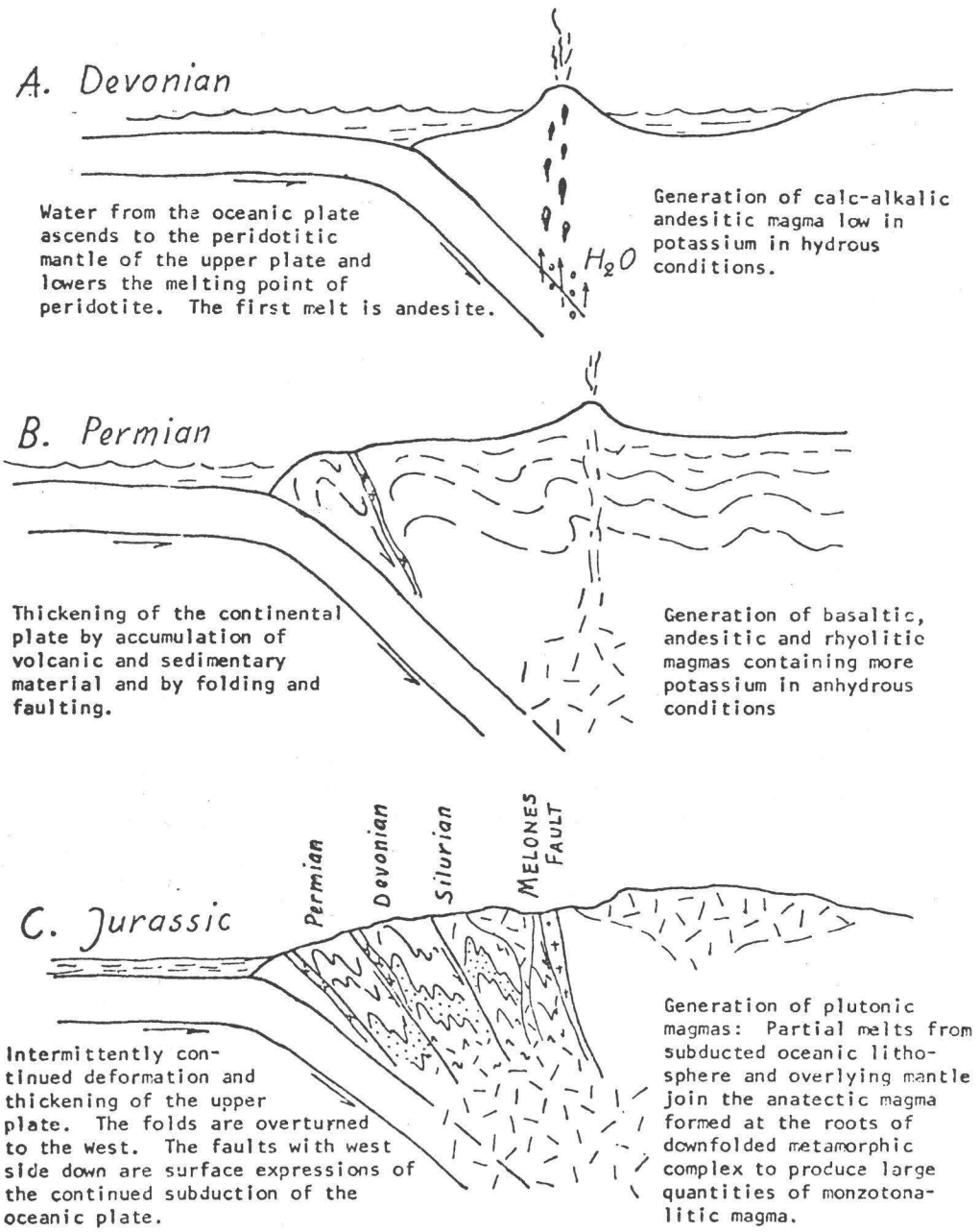

A. Devonian
Water from the oceanic plate ascends to the peridotitic mantle of the upper plate and lowers the melting point of peridotite. The first melt is andesite.
H_2O
Generation of calc-alkalic andesitic magma low in potassium in hydrous conditions.
B. Permian
Thickening of the continental plate by accumulation of volcanic and sedimentary material and by folding and faulting.
Generation of basaltic, andesitic and rhyolitic magmas containing more potassium in anhydrous conditions
C. Jurassic
Permian
Devonian
Silurian
MELONES FAULT
Intermittently continued deformation and thickening of the upper plate. The folds are overturned to the west. The faults with west side down are surface expressions of the continued subduction of the oceanic plate.
Generation of plutonic magmas: Partial melts from subducted oceanic lithosphere and overlying mantle join the anatectic magma formed at the roots of downfolded metamorphic complex to produce large quantities of monzotonalitic magma.

Figure 2. Sketches suggesting structural and magmatic evolution of the postulated subduction zone in the northern Sierra Nevada.

Figure 4

Hypothetical cross section of a continental margin showing zones of magma production (after Hyndman, 1972)

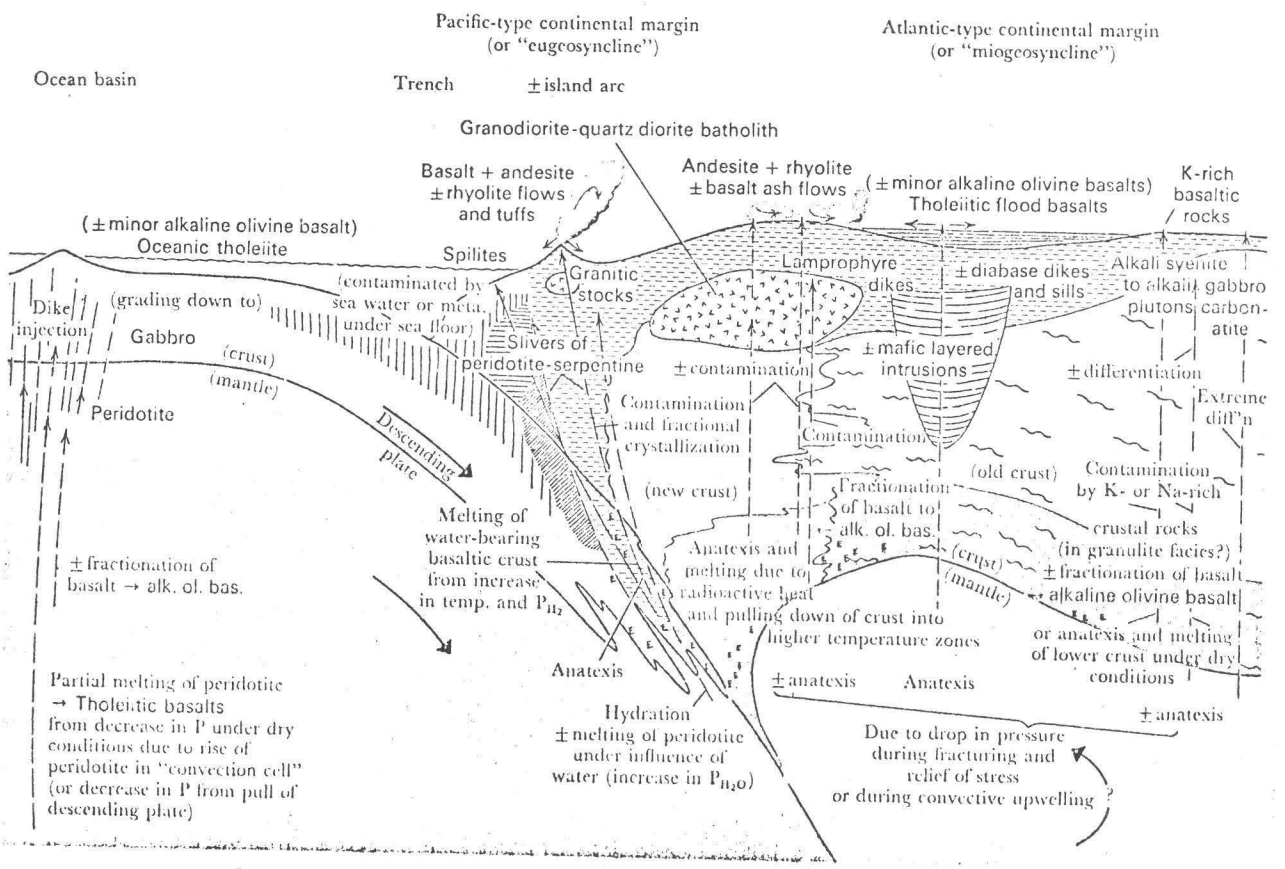


Figure 5

2. Zone melting of continental crust (granitic composition).

3. Anatexis (melting) of eugeosynclinal (basaltic) or miogeosynclinal (granitic) sediments.

4. Any combination of the above mechanisms.

If granitic batholiths are formed strictly from magmas generated by mantle differentiation, they should have low Sr86/Sr87 ratios but if they are derived from zone melting of continental crust, they should have high Sr86/Sr87 ratios. Sediments deposited in eugeosynclines (graywackes) have low Sr86/Sr87 ratios because they are derived from volcanic sources (Island arcs) while miogeosynclinal sediments have high Sr86/Sr87 ratios because they are derived from continental (granitic) sources.

Kistler and Peterman (1973) have shown that the Sierra Nevada rocks have generally middle range Sr86/Sr87 values but that some granitic rocks have anomalously low ratios. Their general conclusions were that both mantle material and lower continental crustal material were combined in the original granitic magma but that some of the Sierra Nevada plutons may be more derived from mantle differentiation than from fusion of continental crustal material. They also discount magma generation from subducted basalts or eugeosynclinal sediments. It seems possible that the quartz diorite zone could represent the mantle derived material while the more felsic, eastern rocks could be representative of the continental crustal fusion process.

For years one of the major problems in igneous petrology has been the "room problem." How could huge quantities of granitic material be produced to form large batholiths which were assumed to extend to great depths in the crust and how could the batholiths displace other material to find room for emplacement. In 1968 Hamilton and Myers published a paper which presented evidence showing that batholiths were essentially flat sheets that were emplaced below a cover of their own volcanic ejecta. Figure 7 shows Hamilton and Myer's idea. More recent work by Hamilton and Myers (1974) has shown that the Boulder batholith in western Montana conforms to their model of a batholith which they describe as being a relatively thin sheet covered by volcanic rocks (produced from the granitic magma itself) and floored by a migmatite/gneiss complex or metamorphosed sediments. This model effectively removes the room problem.

Chase and Talbott (1973) showed that the Idaho batholith was rather passively emplaced in western Montana. Bennett (1973) showed that a section of the batholith was passively emplaced and that the original magma was undersaturated in volatiles, however, late stage differentiates of the original magma became volatile charged and this magma was forcibly emplaced with the result that several different types of plutons were produced from a single magma.

The "Quartz Diorite Line" (Moore, 1958)

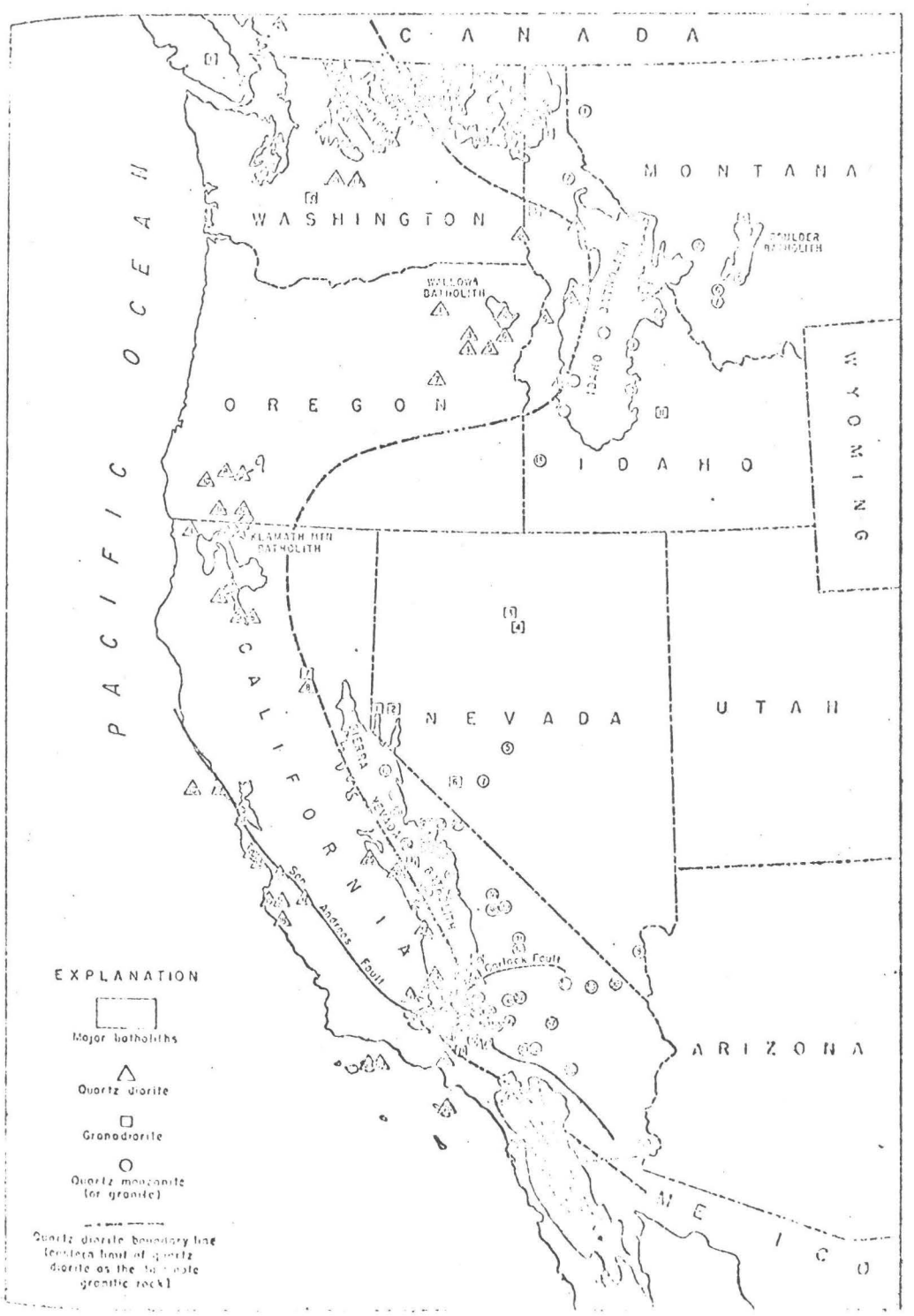

C A N A D A
M O N T A N A
W A S H I N G T O N
O R E G O N
I D A H O
W Y O M I N G
N E V A D A
U T A H
A R I Z O N A
M E X I C O
PACIFIC OCEAN
CALIFORNIA
BOULDER BATHOLITH
WALLOWA BATHOLITH
KLAMATH MTN BATHOLITH
SIERRA NEVADA
San Andreas Fault
Garlock Fault
EXPLANATION
Major batholiths
Quartz diorite
Granodiorite
Quartz monzonite (or granite)
Quartz diorite boundary line (encloses bout of quartz diorite as the dominate granitic rock)

Figure 6

Diagrammatic longitudinal cross section through the Boulder batholith, Montana (Hamilton and Myers, 1968)

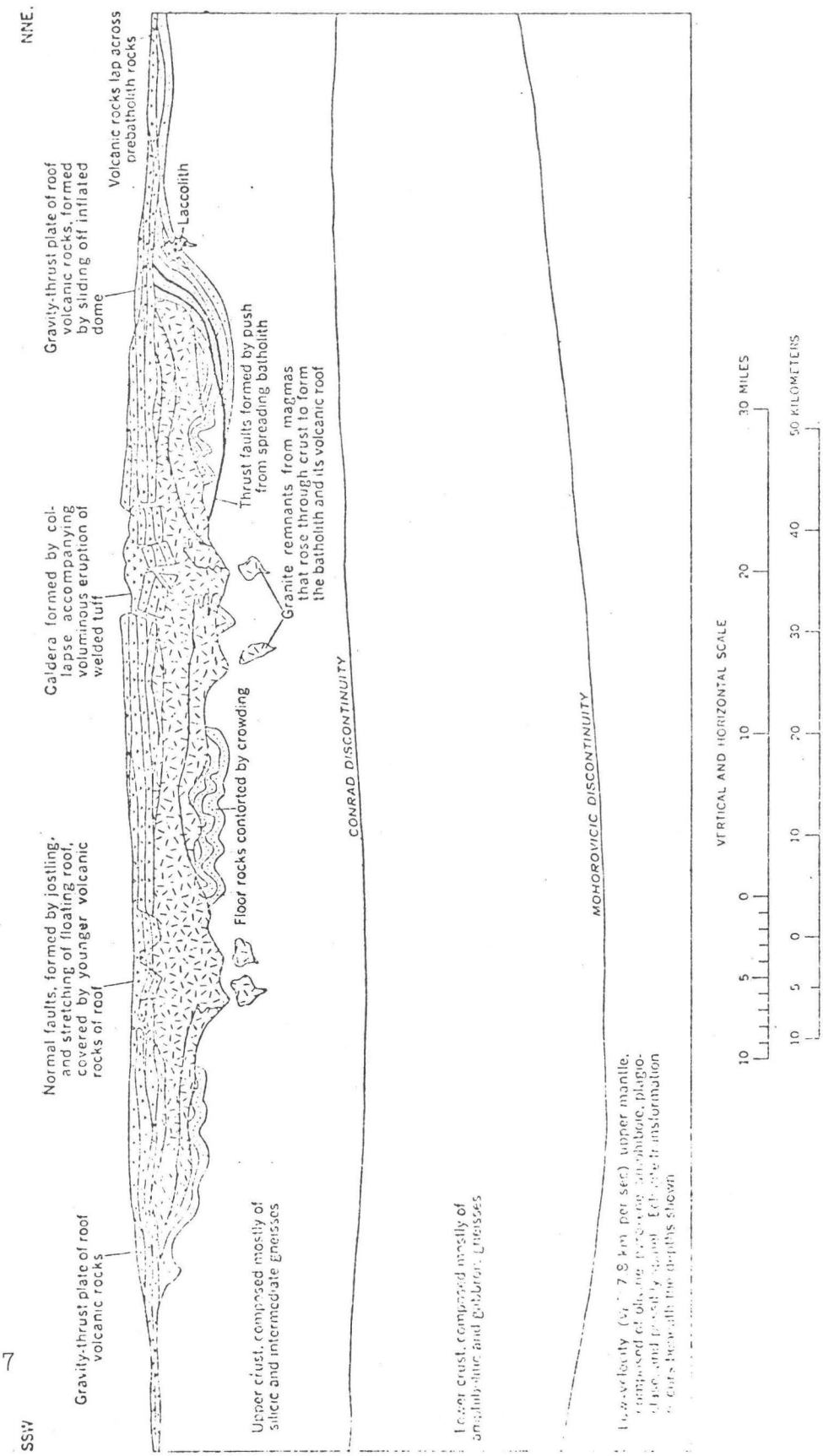

SSW
NNE.
Gravity-thrust plate of roof volcanic rocks
Upper crust, composed mostly of silicic and intermediate gneisses
Normal faults, formed by jostling, and stretching of floating roof, covered by younger volcanic rocks of roof
Floor rocks contorted by crowding
Caldera formed by collapse accompanying voluminous eruption of welded tuff
Gravity-thrust plate of roof volcanic rocks, formed by sliding off inflated dome
Volcanic rocks lap across prebatholith rocks
Laccolith
Thrust faults formed by push from spreading batholith
Granite remnants from magmas that rose through crust to form the batholith and its volcanic roof
CONRAD DISCONTINUITY
MOHOROVICIC DISCONTINUITY
VERTICAL AND HORIZONTAL SCALE
10 5 0 10 20 30 40 50 KILOMETERS
10 5 0 10 20 30 MILES

FIGURE 1.—Diagrammatic longitudinal geologic and crustal section through the Boulder batholith, Montana. The batholith and its roof are of Late Cretaceous age, and the stratified rocks range in age from late Precambrian to early Late Cretaceous.

Figure 7

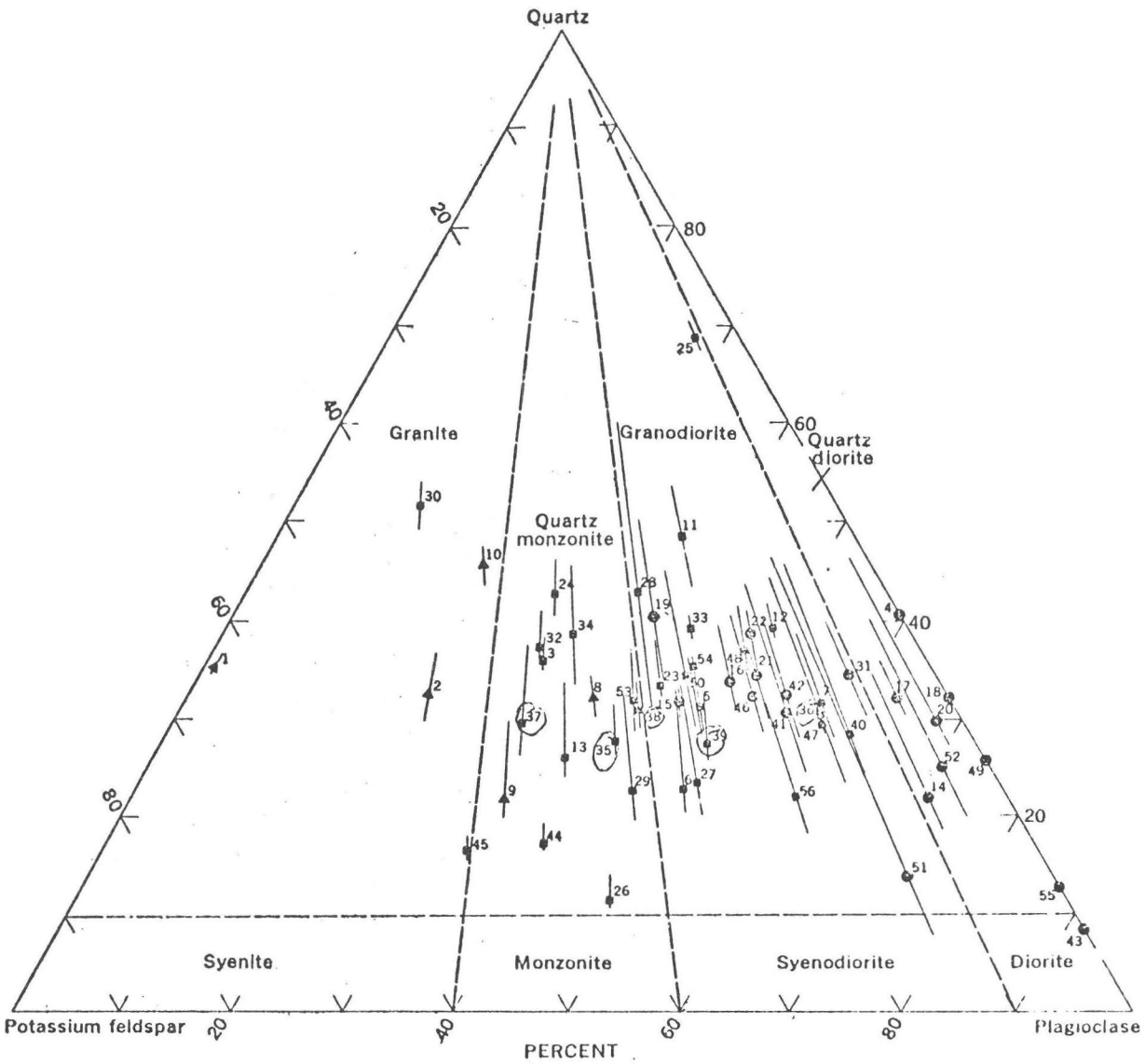
In summary, the Idaho batholith may have formed in the following way. The original batholith magma was produced in an active subduction zone. The source of this magma could have been differentiated mantle material or fused lower continental crust or a combination of these. Anatexis of eugeosynclinal sediments may have also produced part of the magma. As the magma was generated, it rose through the crust probably as individual pods which when reaching shallow depths coalesced and spread out as a thin, sheetlike batholith. This sheetlike batholith was passively emplaced but as the magma differentiated in place new magmas were formed which were volatile charged and this magma was then forcibly emplaced. The Idaho batholith essentially fits the Hamilton-Myers' model. More evidence of this will be presented in the section-subsection discussions.

DISTRIBUTION OF ROCK TYPES IN THE IDAHO BATHOLITH

As was shown in the previous sections, the Idaho batholith is zoned spatially and chemically. More mafic, older rocks lie at the western edge of the batholith while the rocks are progressively more felsic and younger eastwards. Figures 8, 9, and 10 show the composition of the Idaho batholith based on normative and modal analyses. Figures 8 and 9 are based on modal analyses (point counting the minerals on stained rock slabs) while Figure 10 is based on normative analysis (reconstructing a possible mineralogy from the bulk rock chemistry). These diagrams show that most of the batholith is composed of granodiorite and/or quartz monzonite with subordinate amounts of quartz diorite and true granite. In Figure 10 a group of chemically distinct rocks called trondhjemites (Hamilton, 1962) are defined. Trondhjemites are found near the western border zone of the batholith and will be discussed later. Figure 8 is Ross's (1963) modal analysis diagram for the entire batholith, whereas Figure 9 is Bennett's 1973 diagram for a 157 square mile portion of the batholith. Both diagrams show about the same frequency of rock distribution. This illustrates the problems encountered with classification of the rocks in the batholith as a small area may have the same rock types in the same quantity as the entire batholith.

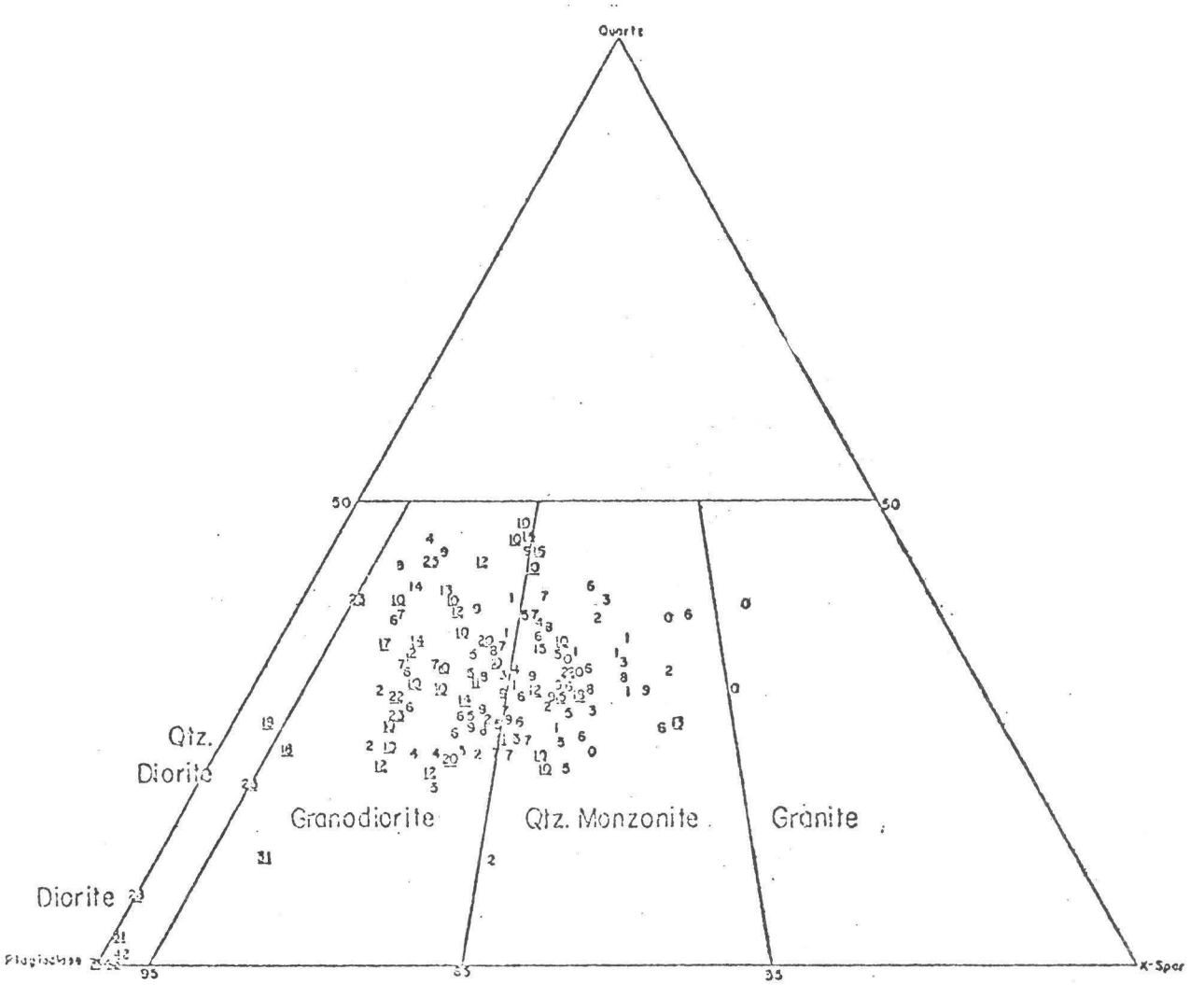
Schmidt (1964) was able to define a sequence of rock types in the batholith near Cascade, Idaho. He distinguished a migmatite border zone, a quartz diorite gneiss zone, a leucocratic quartz diorite zone, a granodiorite zone and finally a quartz monzonite zone. The sequence extends from the western border zone eastwards to the center of the batholith.

Larsen and Schmidt (1958) in a road reconnaissance of the entire batholith (excepting the primitive areas) found the distribution of rocks which is reproduced in Table 1. The distribution of these various units is more appropriately discussed in the sections and subsections portion.



Modal analysis of rocks from the Idaho batholith (Ross, 1963)

Figure 8



Modal distribution of quartz, K-feldspar, and plagioclase in the Idaho batholith (157 square mile section) and the White Cloud stock (Bennett, 1973)

Figure 9

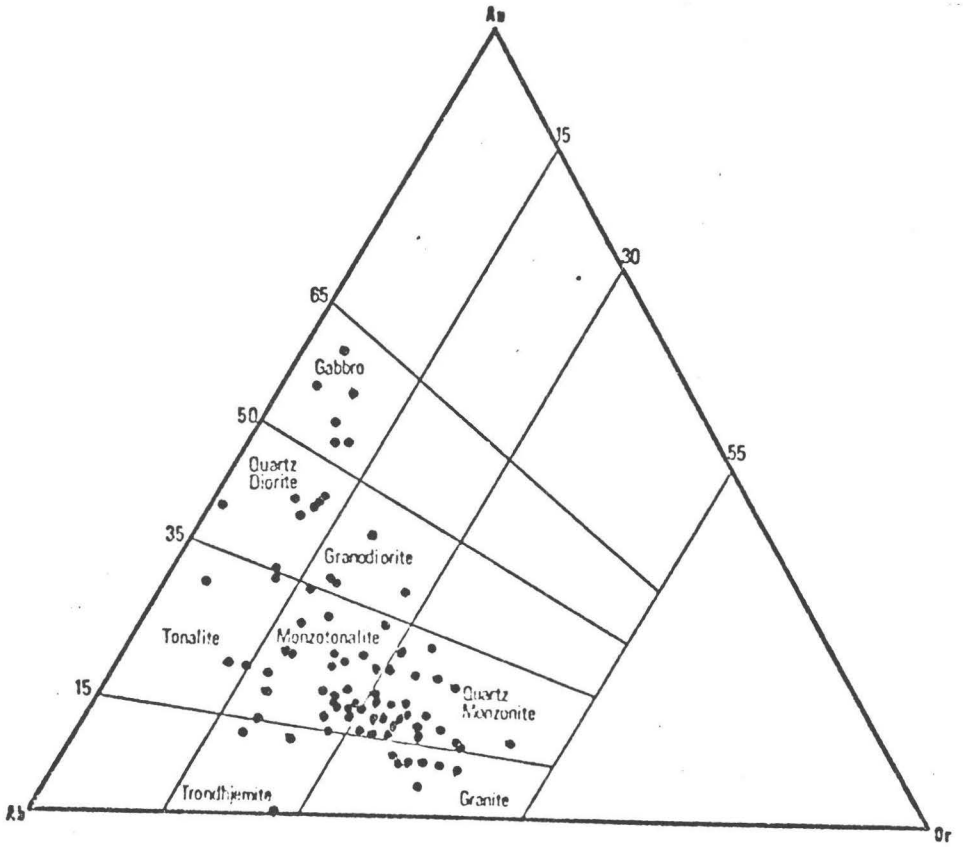


Figure 10. Chemical classification of rocks from the Idaho batholith based on normative analysis (Hietanen, 1963)

Rock type	Percent of batholith	Estimated area (sq mi)	Remarks
Quartz gabbro	<1.0	40	Northeast of Grangeville.
Tonalite	9.0	1,500	Chiefly in north half of batholith.
Granodiorite:			
Cascade type	11.0	1,800	South of Cascade.
Atlanta type	12.5	2,000	Boise River.
Fine-textured quartz monzonite.	25.0	4,000	Chiefly in eastern part of batholith. In part belongs to younger batholith.
Fine dark quartz monzonite.	2.0	300	Elk City area.
Coarse quartz monzonite	37.5	6,000	Chiefly in southwestern part of batholith.
Microgranite	<1.0	40	Chiefly in northeastern part of batholith.
Porphyroblastic gneiss	2.5	400	Chiefly in north half of batholith.

Table 1. Areas underlain by subdivisions of the Idaho batholith after Larsen and Schmidt (1958)

Endomorphism of the Idaho Batholith

Anderson (1942) described what he believed to be evidence of late stage endormorphism of the batholith. He noted that several minerals appeared to have formed or altered after the main part of the batholith had crystallized. The most reliable of his observations concerned large K-feldspar phenocrysts which are frequently observed in the batholith. The phenocrysts appear to be a late stage growth product due to probable potassium metasomatism. With the observed movement of volatiles in the batholith it seems reasonable to assume that at least part of the growth of these phenocrysts was due to late stage metasomatism.

Tertiary Intrusives

A series of plutons outcrop within the area of the Idaho batholith that are intrusive into the batholith and which are dated Tertiary in age. These plutons include the Sawtooth batholith, the Casto pluton, the Crags pluton, and several unnamed plutons in the Selway-Bitterroot area. With the exception of the Crags pluton (gray quartz monzonite) the rest of these bodies are generally pink, leucocratic, quartz monzonite. This pink color makes them quite distinctive in the field. The relationship of these Tertiary plutons to the Idaho batholith (Late Cretaceous) is not known; however, one of the Tertiary plutons, the Sawtooth batholith, contains a very large zenolith of schist included in it. The Idaho batholith in this area has very few zenoliths of any kind. It seems probable that the batholith if emplaced as a sheet would contain few zenoliths, but later Tertiary intrusions which penetrated the sheet could bring up basement zenoliths from below.

Associated with the Tertiary plutons are intrusive dike swarms. These dikes vary compositionally from lamprophyres (ultramafic) to granophyres (granitic). The bulk seems to be latites or andesites which are porphyritic. The relationship of the dikes and the plutons will be discussed later; however, it should be noted that these dikes are reported from widespread areas of the province and must collectively represent a considerable quantity of magma.

THE GENERAL TECTONIC FRAMEWORK OF THE ROCKY MOUNTAIN PROVINCE

Figure 11 is a generalized tectonic map of Idaho. Several important features are noted on the map. The approximate trend of the Snake River downwarp and the parallel trend of the compensatory uplift north of the river plain are very important as much of the drainage pattern within the state is controlled by these features. The tendency of faults on the western border of Idaho to strike north-south has controlled valley development and drainage on this side of the state. The northwest trend of faults and structure in general on the eastern side of Idaho has controlled the orientation of not only drainages but the formation of several mountain ranges including the Bitterroot, Beaverhead, and Lemhi Ranges.

In the northern part of the state the Trans Idaho Discontinuity is partially shown. This structural province may have controlled much of the geology and geomorphology of northern Idaho.

Finally the location of several batholiths within the state are noted. These large masses of rock are responsible for many of the above mentioned structural features.

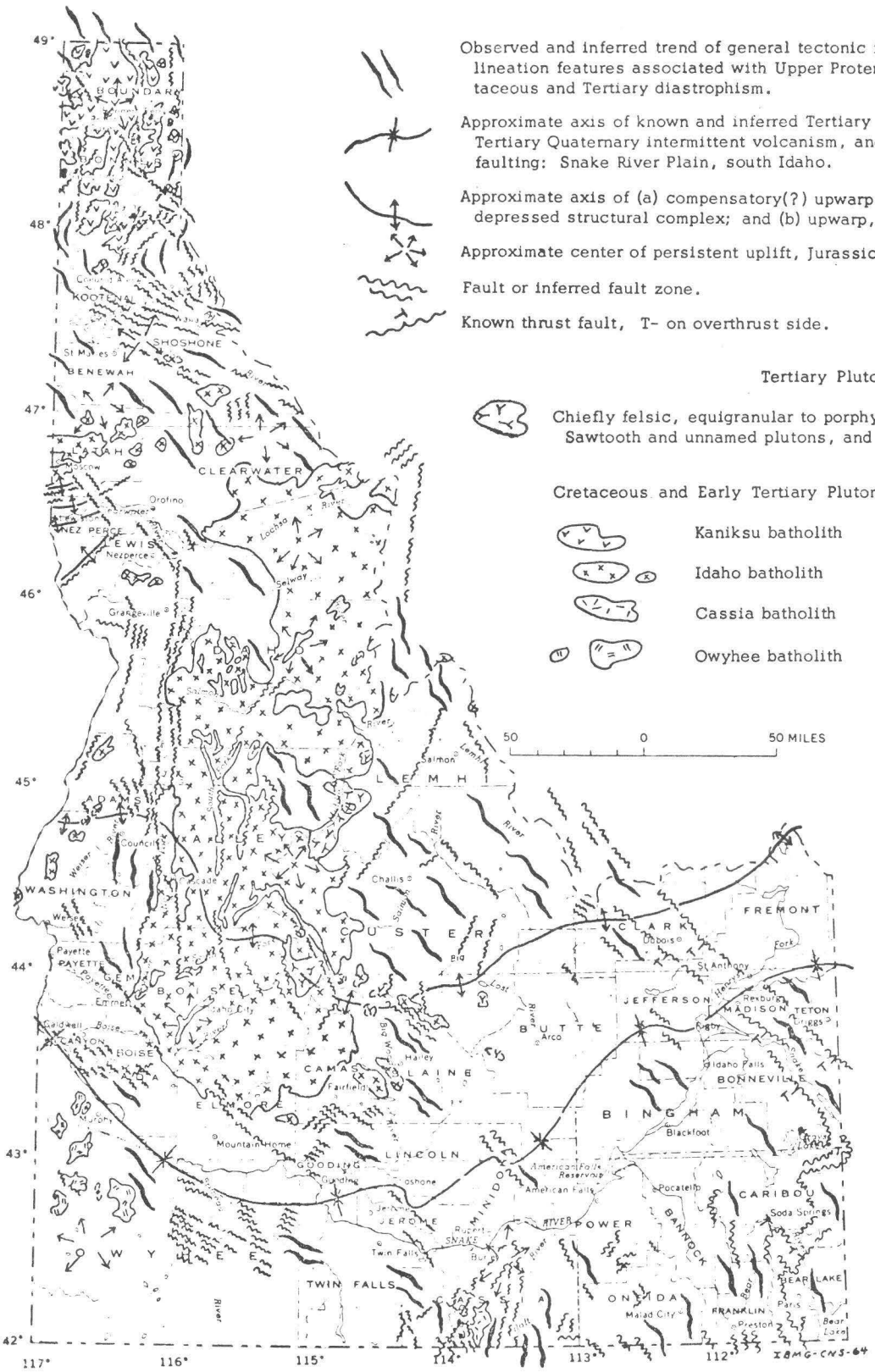


Figure 11. Preliminary Tectonic map of Idaho (Savage, 1964)

THE GEOLOGY OF SECTION I

THE SOUTHERN BATHOLITH SECTION

Parts of the Hailey, Challis, Baker, and Boise AMS maps are covered by Section 7. The Section roughly includes the area from 114° 15' to 116° 15' west longitude and from 43° to 45° north latitude. Major drainages in the section are, in the eastern part, the headwaters of the Salmon, and in the northern part, the Middle Fork of the Salmon with its tributaries, Big Creek and most of Camas Creek. The northwest portion of Section I is drained by the South Fork of the Salmon, part of the Payette River and its tributaries, the North Fork, Middle Fork, South Fork, and Dead Wood River. The southwest part of the Section is drained by a part of the Boise River system.

Almost the entire area is underlain by Idaho batholith with overlying patches of Tertiary Challis volcanics and a few areas of Tertiary-Quaternary sediments. Section I is primarily a study of different granitic rock types. The topography in the Section is characterized by steep relief and is quite rugged like most of the Rocky Mountain Province. Elevations range from about 4,000 feet near the Snake River Plain to over 9,700 feet in the Trinity Mountains and over 10,000 feet in the Smokey Mountains.

Subsection 1 - Smith's Prairie and Foothills (complex) Subsection

Subsection 1 is a foothills Subsection which is drained to the west by the Boise River system and to the east by the Camas Creek system. Relief ranges from 4,800 feet to over 6,000 feet. Most of the area is underlain by Idaho batholith which is probably "Atlanta type" granodiorite (Larsen and Schmidt, 1958). Subsection 1 is characterized by a curvilinear pattern (see Hailey linear map) which is fully discussed in the next subsection.

Subsection 2 - Trinity-Soldier Mountain Foothills (complex) Subsection

Subsection 3 - Little Fork-Boise River Canyon and Streamcut Land Subsection

Subsection 4 - South Fork Boise River Uplands Subsection

The only information which is available in Subsections 2, 3, and 4 is from the road log of Larsen and Schmidt (1958). However, just on the basis of this log a major change in geology is indicated within these subsections.

A little south of Arrow Rock Reservoir (Subsection 3) is a small band of coarse quartz monzonite (Ki_3ch). This same quartz monzonite is found in the central area of Subsection 4 and a small tip of the same rock type is noted just north of the Anderson Ranch Reservoir in Subsection 2. These outcrops of quartz monzonite define a central area of this rock type which is completely surrounded by granodiorite (Ki_3be) which Larsen and Schmidt call the "Atlanta type" granodiorite.

Examination of the linear map for the Hailey AMS map shows a radial pattern of linears which are centered on Steel Mountain (just below the first "o" in Sawtooth Mountains). These linears trend northeast, north-northeast, southwest and southeast. Careful examination of the smaller linears at the outer edge of these four radial linears reveals a strong concentric pattern. The concentric pattern surrounds the four radial linears and includes most of the area mapped as quartz monzonite. This concentric pattern continues outward from this center on a larger scale, with a very pronounced curvilinear trend; starting in the eastern half of Subsection 2, the linears trend east-northeast, in Subsection 5 the linears trend north-northeast and in Subsections 6 and 7 the trend is north-northwest and northeast. On the western side of the Hailey AMS sheet the linears in Subsection 10 trend north-northeast then the pattern changes to northwest in the northwestern portion of Subsection 3 and the trend is finally east-west in the western portion of Subsection 2. It is very possible that the radial linear pattern in the quartz monzonite which is surrounded by a concentric ring pattern in the Atlanta granodiorite represents a regional ring fracture system. The quartz monzonite is probably a late stage batholith differentiate which intruded the granodiorite (which is the sequence of intrusion observed in other areas of the batholith). The intrusion would have produced the observed linear-fracture patterns.

The above observations are in part speculative and have never been field checked or described before. However, if these observations were supported by field work, almost all of the structures in the northcentral and northwestern section of the Hailey AMS map have been controlled by these events.

Some topographic control may be represented by the different rock types. The Trinity Mountains are composed of quartz monzonite and are higher and less dissected than the Atlanta granodiorite to the southwest of the Middle Fork of the Boise River in Subsection 3. This may in part be due to differences in composition, jointing, or some other structural control.

Larsen and Schmidt (1958) described the Atlanta granodiorite and the coarse quartz monzonite as follows.

Granodiorite of the Atlanta Area

> "A very large body of rather uniform rock of moderately dark color and chiefly a granodiorite but in part a quartz monzonite underlies most of the drainage of the South and Middle Forks of the Boise River. A body of the light-colored coarse-grained quartz monzonite occupies the central part of this area. This granodiorite was observed for a distance of 70 miles in a east-west direction, and it probably extends further, for it is covered by basalt on the west border and was not followed to its east limits. In a north-

south direction it has a maximum width of 40 miles, with the quartz monzonite body in its central part. It occupies an area of more than 2,000 square miles. Many local variations in the mass and numerous small inclusions are present, but the main rock from the western exposure, about 10 miles southeast of Boise, up to the Middle Fork of the Boise River and up to and beyound Atlanta for a distance of over 55 miles is rather uniform. Along the other traverses the rock is similar.

The rock is seriate porphyritic with crystals of plagioclase that range in length from 4 millimeters to very much shorter. The interstitial material is small in amount and is made up of grains as small as 0.1 millimeter across. The plagioclase is rather strongly zoned, with cores of andesine and broad borders of albite, and has an average composition of about An_{32}. The fine material is chiefly quartz and microcline, in part intergrown. The rocks of the traverse along the Middle Fork of the Boise River are chiefly intermediate between quartz monzonite and granodiorite, contain little or no hornblende, and have a moderate amount of accessory minerals. Those to the south are chiefly granodiorites and contain hornblende as well as biotite. They have abundant large grains of sphene, allanite, monazite, magnetite, and apatite and contain some secondary chlorite, epidote, and minor sericite. The rocks are weathered near the surface, but they have undergone less hydrothermal or deuteric alteration than most of the rocks of the Idaho batholith."

Coarse Quartz Monzonite

"A nearly white coarse-grained quartz monzonite is the most widespread and abundant rock of the batholith. It constitutes 32 percent of the rocks traversed, and its actual extent may be somewhat greater as only a few short traverses were made in the central part of the batholith where this type of rock is believed to predominate.

The rock is light gray to nearly white and is coarse grained. Most of the rocks are porphyritic or scriate porphyritic, with an average grain size of about 2 millimeters, and they are fairly uniform over wide areas, although there is some range in texture, expecially in the size of phenocrysts. Feldspar crystals more than a centimeter long are present in most of the rocks, and in some rocks they are more than 5 centimeters long. They are

commonly flesh colored, white, or clear. The matrix feldspars are white. Biotite is present in small amounts, and large muscovite plates are present in many of the rocks.

Under the microscope the feldspar phenocrysts are seen to be chiefly orthoclase (untwinned), with some streaks of microcline along fractures. A small amount of albite is also present in very thin lamellae. The large orthoclase crystals tend to mold about and enclose quartz and plagioclase, showing that they are late crystals. They are intergrown with quartz along their borders and locally are partly replaced by myrkemite.

The plagioclase crystals are zoned with some recurrent zones, but they are chiefly cores of An_{25}, broad outer zones of An_{17}, and a skin of albite, which is best developed adjacent to orthoclase.

The biotite is partly altered to chlorite. There is secondary sericite and a few larger plates of muscovite that are probably primary. The rocks are all somewhat altered and contain secondary sericite, rare calcite, epidote, and chlorite.

Most of the rocks have nearly equal amounts of orthoclase and plagioclase, about 30 percent quartz, and a few percent biotite. Some rocks have nearly as much muscovite as biotite, others have no primary muscovite. Magnetite, zircon, monazite, apatite, allanite, and, rarely, sphene are the accessory minerals recognized, and in some specimens they are present in crystals 0.5 millimeter across. The muscovite-bearing rocks contain small amounts of garnet, xenotime, and fluorite. The accessory minerals are present in very small amounts -- on the basis of heavy-liquid separations from large samples of rock, they appear to make up less than 1 part in 10,000 of the rock and probably no more than 1 part in 100,000.

In some places, for distances of a few hundred feet inward from the contacts with adjoining rocks, the coarse muscovite-bearing quartz monzonites contain abundant small (1-2mm across) miarolitic cavities lined with well-formed crystals of quartz and orthoclase." (Larsen and Schmidt, 1958)

Subsection 5 - Upper South Fork Boise River Streamcut Lands Subsection

Subsection 6 - Smoky Mountain Trough Lands Subsection

There is very little data reported from Subsection 5 and 6. A portion of Larsen and Schmidt (1958) road log extends through Subsection 5 and shows Atlanta type granodiorite as the rock type. There are some small outcrops of the Challis volcanics and Wood River Formation in the southeastern part of Subsection 5. Ross (1956) gave a short report on the southwestern part of Subsection 5 but provided few details on structure.

Subsection 7 - Montezuma Fault Lands Subsection

Subsection 7 contains the Atlanta Mining District which was examined by A. L. Anderson (1939). Anderson described the most prominent features of the area starting with the Atlanta Basin (northern part of 7). The east wall of this valley is the Montezuma fault scarp formed by a fault which separates the Atlanta district from the much higher Sawtooth Range to the east. The area is drained by the East Fork of the Boise River. To the south of the basin is the Atlanta Hill which was the location of most of the mines which were active in the late 1800's.

Anderson said that most of the area is composed of quartz monzonite which is porphyritic in places and prominently porphyritic in the Atlanta Hill area. Granodiorite and quartz diorite crop out at the eastern edge of the Subsection and these rock types represent the roof facies of the batholith. Aplite and pegmatite dikes are numerous throughout the region but are more concentrated in the Atlanta Hill area. Anderson described a Tertiary dacite/granophyre dike swarm which is about three miles wide and is continuous for more than 11 miles from where it crosses the Boise River several miles below Atlanta. The belt extends in an east-northeasterly direction along the northern side of the river and goes over Graylock Peak. The dike swarm is similar to one described later, by Anderson (1943a) in the Boise Basin and to a porphyry complex described by Ross (1934c) in the Casto quadrangle.

Older surficial glacial deposits cap many of the ridges around the Atlanta basin while younger material fills the basin. Anderson described a series of Alpine glacial morraines which extend over the basin.

The Atlanta Hill area is a shear zone with a structural trend of N 50° to 75° E and a secondary trend of N 60° to 85° W. Numerous aplite and pegmatite dikes within the Hill have the same trends. Anderson believed that the disturbance which formed the shear zone occurred in the Tertiary and therefore was not associated with the batholith emplacement. The Tertiary dike swarm has the same strike as the shear zone, N 60° to 70° E but has a different style

of faulting than the shear zone. Anderson said that the pegmatites and aplites in the Hill were probably not emplaced at the same time as the dike swarm. Within the batholith itself are two prominent fault sets, one trending N 50° to 60° E (parallel to aplite and pegmatites) and another trending west-northwest.

Anderson noted that the Montezuma fault (the major structural feature in the area) trends northwest and is first observable on the Hailey AMS map slightly northeast of Atlanta. This fault is Tertiary in age as it offsets the dike swarm. Most of the movement on the fault was vertical (2,000 to 4,000 feet) whereas the shear zone in Atlanta Hill formed by horizontal movement.

Subsection 8 - Swatooth Foothills Subsection

Reid (1963) mapped the central and southern portion of the eastern half of Subsection 8. The Subsection contains the Montezuma fault which is the prominent geological feature. Reid said that east of a north-south line drawn through Atlanta, the Idaho batholith is quartz diorite or granodiorite in composition while west of this line the rock is granodiorite or quartz monzonite. Schmidt and Larsen's (1958) road log through the area shows Atlanta type granodiorite (Ki_3be) in the southern portion of 8 and coarse quartz monzonite (Ki_3ch) in the northern part. Reid mapped several igneous bodies which he described as leucocratic quartz monzonite and which were intrusive into the batholith granodiorite. This leucocratic quartz monzonite was described as follows:

Leucocratic (white) Quartz Monzonite and Related Rocks

> "The mineralogy of the leucocratic (white) quartz monzonite that intrudes the Idaho batholith is similar to that of the Idaho batholith. Essential minerals are quartz, orthoclase, microcline and plagioclase (An_{8-30}). The plagioclase is more sodic than that of the Idaho batholith, which distinction, along with the light color of the rock, constitutes the major differences between this rock and rock of the Idaho batholith. Accessory minerals are biotite, muscovite, zircon, apatite, allanite, magnetite, sphene, and garnet. Alteration products are clay, sericite, epidote, and pennine. The potassium feldspars are somewhat perthitic. The texture is hypidiomorphic granular." (Reid p 18619)

The description of the leucocratic quartz monzonite is very similar to a description of light colored intrusives in the Pioneer Mountains (Dover, 1973) and the White Cloud stock description (Bennett, 1973). Most of these light colored intrusives are probably late stage differentiates of the Idaho batholith.

Subsection 9 - Lowman Uplands Subsection

Subsection 9 contains the headwaters of Moore's Creek which is a major tributary to the Boise River. Larsen and Schmidt's (1958)

road log is the only data available in the area and it shows coarse quartz monzonite (Ki_3ch) as the dominent rock type. The quartz monzonite represents a change in rock type from the Atlanta granodiorite to the south in Subsection 3. The contact between these two rock types may exert some structural control in the area. A strong linear lies along Moore's Creek and could coincide with the contact region.

Subsection 10 - Boise Basin Subsection

Anderson (1943a) mapped the central part of Subsection 10 in a study of the mining areas in the Boise Basin. The major geologic feature in the mapped area is a Tertiary dike swarm or "Porphyry Belt" which is located in the northwest and southwest corners of the Hailey and Challis AMS maps respectively. This dike swarm is similar to the one described by Anderson (1939) in the Atlanta district (Section I, Subsection 7) and has the same northeasterly trend. The most common rock type in the swarm is a quartz-hornblende-biotite monzonite porphyry. Larsen and Schmidt (1958) mapped the batholith in Subsection 9 as a coarse quartz monzonite (Ki_3ch).

The east-west trending divide between the Boise River-Moore's Creek drainage in the South and the Payette River system to the north was discussed by Anderson. Because the topography of this subsection and surrounding subsections has been greatly influenced by this divide, Anderson's discussion of the geologic history of the Boise Basin from Late Miocene times follows. Much of this discussion can be applied to the Lost River-Big Wood River drainage to the east (analogous to the Boise Basin) and the Salmon-Pahsimeroi River system to the northeast (analogous to the Payette River system).

> "Apparently as the basaltic extrusions died away near the close of the Miocene, the igneous activity was succeeded by a mild crustal disturbance involving differential uplift to the north. The low, essentially flat surface that characterized the general region then became a tilted plain, sloping gently to the south. The differential uplift renewed the cutting power of the streams and produced marked changes in the drainage patterns. The ancestral Boise and Payette Rivers continued to flow in a westerly direction as at present, parallel to the axis of uplift, but the tributaries on the north sides of the main Boise and Payette Rivers had the advantage of increased gradients and gully cut back far up the slope. These streams, favored by slope, successfully competed against the tributaries of greatly reduced gradients that joined the trunk streams from the south and that had to flow against the slope to maintain their courses. Capture of the drainage basins of the handicapped streams by those favored by surface slope was imminent, and in this way the tributaries of the Boise River robbed the Payette River of most of its drainage area, except on the north, and those joining the Payette from the north worked far up

the slope against the headwater of the Salmon River. A very one-sided drainage pattern was thereby evolved. During these adjustments one of the tributaries of the Boise River captured Grimes Creek water where it entered the Payette River and has maintained this capture to the present day.

Adjustment of the drainage to the tilted slope probably required much of Pliocene time for its completion. During this interval the uplifted surface was carved into a system of low hills and flat-topped ledges separated by relatively shallow valleys, in places as much as several hundred feet deep. The products of erosion were carried westward and probably deposited on the somewhat eroded Payette formation and basalt as beds of the Idaho formation. In the later part of the Pliocene the region stood low, and its surface resembled that of the present upland areas.

The region was then subjected to another crustal disturbance beginning at the close of the Pliocene epoch and continuing intermittently to the present day. Elongated segments of the crust were broken into faults and the segments tilted into block mountains, of which Boise Bulge stands as a conspicuous example. The region was also differentially warped, and such intermontane depressions as Boise Basin and Garden Valley were produced partly as a result of faulting and partly as a consequence of differential uplift. The region as a whole was elevated to the levels at which it stands today.

The general uplift greatly increased th cutting power of the main streams, and the elevated erosion surface was deeply and intricately dissected. The South Fork of the Payette River rapidly deepened its channel, particularly in the high upland east of Garden Valley and across the mountain block to the west. Grimes Creek, because of its distant connection with the Boise River, was not so materially affected by the uplift and continued on as before; but the Payette, in consequence of its steepened gradient produced by the tilting of the faulted crustal segment against the base of Boise Ridge west of Garden Valley carved its canyon far below the level of Grimes Creek and at the same time undercut and moved back the divide, separating the two where the creek now appears to flow for nearly 3 miles in a narrow, one-sided trough high on the canyon wall.

The warping of the low, hilly surface in Boise Basin steepened the gradients of the streams flowing down

> the flanks of the basin, causing them to erode deep valleys and canyons and to carve canyons of even greater depth in order to maintain their courses across the brim on the opposite side. Conditions, locally, as in Boise Basin, favored the accumulation of valuable placers but did not promote deep weathering and most of the weathered material previously formed during the Pliocene, when erosion was not nearly so active, was eroded. After Moore Creek and lower Grimes Creek had carved their canyons nearly to their present depths, basalt poured out from a vent just above the junction of the two streams and flowed southward, eventually reaching and spreading over the margin of the Snake River Plain. The district escaped the Pleistocene glaciation, but degradation was interrupted several times, probably because glaciers at the headwaters of the main rivers freed so much material that the streams could not carry all of it away and alluvium accumulated along the valley bottoms, later to be partly removed as the stream load returned to normal. The mountains thus carved out of the uplifted, faulted and warped Tertiary erosion surface are now in the process of being worn away." (Anderson's discussion)

Anderson (1934) mapped the Pearl-Horseshoe Bend Gold Belt which is located in the central-western part of Subsection 10. This area contains another "Porphyry Belt" similar to the Boise Basin Belt and this dike swarm has the same trend, N 60° E. This Belt was not observed to connect to the Boise Basin Belt. The rest of the geology is similar to the Boise Basin description.

Anderson described a major fault which strikes north and southwest from Horseshoe Bend and has caused the Payette River to meander widely over the Payette Formation (unconsolidated Tertiary sediments). These sediments are on the down dropped block of the fault and fault action has caused the Horseshoe Bend to form in the river. The Payette flows through a narrow canyon to the west which is in more resistant Idaho batholith rocks. This north-south fault along with several other parallel linears is evident on the Boise and Baker linear maps. These linears are responsible for most of the structural basins on the western margin of Idaho.

Subsection 11 - South Fork Payette Canyon and Streamcut Lands Subsection

Subsection 11 has very little information about it other than the road log of Larsen and Schmidt (1958), which classifies the entire area as coarse quartz monzonite (Ki3ch). The author has traveled the road from Lowman west to Garden Valley several times and agrees with Larsen and Schmidt. There are several dike swarms (lamprophyres, latite, andesite) that cut this portion of the batholith along the road. The development of the Payette drainage has been covered in the discussion of Subsection 10.

Subsection 12 - Sawtooth Mountains Subsection

This Subsection includes the Sawtooth Mountains and a small area north of this range (Reid, 1963) map, the Sawtooth. The range is a glaciated horst which is formed between the Montezuma fault on the west and the Sawtooth fault on the east. Most of the range is composed of Sawtooth batholith (TEI_3). This Tertiary intrusive is quite distinct from the Idaho batholith. Reid described the pink quartz monzonite which makes up the batholith as follows:

Sawtooth Batholith--Leucocratic (Pink) Quartz Monzonite and Related Rocks

"The mineralogy of the leucocratic (pink) quartz monzonite that intrudes both major varieties of igneous rock described above is somewhat different from that of the other two plutonic igneous rock units in the Sawtooth area. Essential minerals are quartz, mesoperthitic, orthoclase and plagioclase (An_{2-32}). Minor hornblende, biotite, and muscovite are present in some specimens. Accessory minerals are zircon, apatite, sphene, allanite, ilmenite, magnetite, fluorite, and aquamarine (rare). Alteration products are sericite, leucoxene, pennine, and clay.

The mesoperthitic orthoclase in this rock is estimated to comprise about two-thirds of total feldspar. Many workers would prefer to call such a rock granite. However, because the albite content of the perthite would surely bring the rock chemically into the quartz monzonite class, I prefer to call it quartz monzonite.

Where pink quartz monzonite intrudes older igneous rocks, a zone of pervasive green alteration several hundred yards wide has developed in many places. Thin section study reveals that the altered rock has been subjected to a rather strong, pervasive crushing or non-directional mylonitization. Original plagioclase has been altered to albite, and intensive clay alteration of potassium feldspar has occurred. The green color is caused by sericite-chlorite-epidote aggregates distributed in small crush zones.

Pegmatites of the pink quartz monzonite have the following essential minerals: quartz, mesoperthitic orthoclase, and plagioclase (An_{2-17}). Accessories are magnetite, apatite, zircon, allanite, fluorite, aquamarine, and garnet. It is to be noted that the mineralogy of these pegmatites is closely related to that of the enclosing rock.

> Trends of pegmatite dikes in the pink quartz monzonite are rather varied, but most dikes dip at rather low angles. A few, near-vertical dikes trend to the north, northeast, and northwest. There is little relation between pegmatite dike trends and major joint trends in the pink quartz monzonite. Probably, the joints were formed later than most of the pegmatite dikes." (Reid, 1963)

Reid's description of late stage Idaho batholith intrusives (leucocratic quartz monzonite, Ki_3a) has been given in Section I, Subsection 8. The older Idaho batholith rocks are primarily granodiorite (Ki_2a). A large outcrop of mica schist (mpCl) called the Thompson Peak Formation is included in the Sawtooth batholith and may be a large zenolith. This unit is described by Reid as being similar in appearence to the metamorphosed Wallace Formation to the north.

Two steeply dipping joint sets trending northeast and northwest are responsible for the Sawtooth appearence of the range. Some of the best Alpine glacial features in Idaho can be seen in and from the Stanley Valley where long lateral and terminal morraines have been deposited from the Sawtooths. The range itself contains many cirques, horns, aretes, tarns and other Alpine features.

Subsection 13 - Stanley Basin Subsection

Subsection 13 includes the Stanley Basin. Williams (1961) described the basin as "an elongate intermontane valley in the headwaters of the Salmon River." The basin was most likely formed by the Sawtooth fault. On the western edge of the basin the fault is marked by a steep scarp whereas the eastern border of the basin has more gentle slopes. Tectonic basins like the Stanley Valley are common in Idaho and other basins like the Warren and Salmon Valleys will be described latter.

Two periods of glaciation were recognized in the basin by Williams (1961) which he correlated with the Bull Lake and Pinedale glacial periods (after Blackwelder, 1915). Williams also noted that the Sawtooths had been more severely glaciated than areas to the east of the basin. He believed this was because the Sawtooths have higher elevations than the eastern areas and would therefore be more severely eroded by glacial action.

Williams explained differences in the longitudinal profile of the Salmon River in the Basin by showing that the river flows over sediments which are divided into coarse and fine material. The

river can reach grade in the fine material but cannot in the areas of coarse boulders and cobbles. The areas underlain by coarse material are characterized by rapids while the areas underlain by fine sediments have relatively quiet water.

Subsection 14 - Stanley Uplands Subsection

Subsection 14 covers the area of Idaho batholith northwest of and paralleling the Stanley Basin. Major drainages in the region are the Salmon River (Valley Creek). Several areas just north of the Salmon River are covered by Challis volcanics (Latite-andesite member, Tee2 and Germer tuffaceous member, Tov). Bennett (1973) described the pie-shaped section of Idaho batholith just east of Stanley. The rock type in the area is primarily granodiorite at elevations above 7,000 ft. and quartz monzonite below 7,000 ft. Pegmatites and aplites are localized in the lower quartz monzonite section. The vertical zonation is interpretated as differentiation in place by the original batholith magma.

Faulting in the Subsection was described by Choate (1962) who noted three major fault sets, a N 70° W set (primarily shear), a N 30° E set (shear and tension) and a N 65° E set (shear). Other fault sets were considered to be complements of these three. Choate described jointing in the area as follows;

Joints

> "Joints are found in all rocks of the Stanley area, but are best developed in the granitic rocks. Joints in granitic rocks probably developed in part from cooling stresses and in part from deformational (both shearing and tensional) stresses. At any one locality in granitic rocks at least two systems of joints are present; however, one system is generally dominant. Most joint systems are composed of three subperpendicular joint sets; in many cases two sets of a system dominate over the third set. Distances between fractures in any set range from 1 to 4 feet. Attitudes of joint systems are not constant throughout the area.
>
> Most of the thicker dikes contain a dominant joint system consisting of three sets of cooling fractures. One set is parallel to dike walls; the other two sets are generally perpendicular to dike walls and to each other. In some cases, cooling fractures aid in determining dike attitudes." (Choate 1962)

The igneous rocks in the Subsection were described by Choate as being coarse grained, porphyritic, granitics in the east which graded to fine to medium grained granitics in the west. A Tertiary quartz latite-dacite porphyry complex was described near Stanley Creek and is similar to other dike swarms described previously. Choate mentioned terrace evidence of two stages of glaciation in the area. (See Williams' discussion in the previous Subsection.)

The linear map of the Challis AMS sheet shows strong linears along the Salmon River, Valley Creek, and most of the other drainages in the area. This is typical of the batholith drainage system as most of the major streams are probably fault controlled.

Subsection 15 - Bear Valley-Landmark Basin and Uplands Subsection

Subsection 15 is underlain by an extensive portion of the Idaho batholith which, on the basis of Larsen and Schmidt's (1958) road log, is mostly coarse quartz monzonite with a small stretch of granodiorite in the northern part of the area. The Subsection is drained by the South Fork of the Salmon to the north and the South Fork of the Payette River to the south. The major features in 15 are the two valleys, Bear Valley and Landmark Basin.

Schmidt and Mackin (1970) discussed the geology of Long Valley (just west of Cascade and outside the map area) and Bear Valley in the southeastern part of Subsection 15. They point out that the north-south oriented basins and valleys are probably fault controlled as was postulated by Anderson (1943a). A basin and range type of topography would develop from these faults and downthrown blocks would become valleys which would be filled with fluvial and glacial material.

The changes in the drainage of the Payette River system are related to Tertiary uplift and subsidence as discussed by Anderson (1947). Schmidt and Mackin were looking at heavy mineral placers in the glacial and fluvial material which fills Bear Valley and Landmark Basin and covers much of the underlaying Idaho batholith. An older, pre-glacial series of lacustrine and fluvial deposits were later covered by two stages of glacial deposits assigned to the Bull Lake (oldest) and Pinedale (youngest) glacial periods of Blackwelder (1915). Schmidt and Mackin describe these deposits in Bear Valley as follows.

Bull Lake and Pinedale Glacial Deposits

> "In contrast to the little-modified depositional roughness of the Pinedale moraines, Bull Lake moraines are distinguished by a smooth hummocky topography, clearly morainal but subdued by erosion. Closed depressions which characterize Pinedale moraines are generally drained or filled on Bull Lake moraines. Most of the depositional surface on Bull Lake outwash has been destroyed, whereas most of the Pinedale surface remains. The contrast in degree of weathering is conspicuous; for example, in Bull Lake deposits a large percentage of granitic pebbles and boulders disintegrates readily whereas in Pinedale deposits under otherwise comparable conditions a large percentage is fresh. The Bull Lake terminal moraines are commonly downvalley from the

Pinedale moraines, the distance between them ranging from less than 1 mile to 4 miles, depending on topographic relations.

Bull Lake and Pinedale moraines are ideally displayed at the north end of Long Valley, where patterns of compound piedmont lobes are defined by morainal loops; the Payette Lakes lie within the Pinedale moraines, which in turn are surrounded by outer loops of Bull Lake age. South of the Bull Lake moraines, the dissected Bull Lake outwash plain is partly buried beneath a smooth-surfaced plain of Pinedale outwash. Much of the Pinedale outwash fills broad valleys cut in the Bull Lake outwash during the interglacial interval.

Short steep glaciers from West Mountain barely reached the valley floor. Although morainal forms are not well developed, Bull Lake and Pinedale moraines and outwash are readily distinguished in most places. Both consist mostly of subangular metamorphic rock types and basalt that contrast sharply with the exotic quartz-pebble gravel mentioned earlier.

Big Meadow is a broad north-south segment of the upper part of Bear Valley Creek (pl. 2, T. 11 N., R. 8 E.) and is bordered on the west by a range high enough to have supported Bull Lake and Pinedale glaciers which extended to the valley floor. Big Meadow is bordered on the east by a relatively low ridge that was not glaciated during the late Pleistocene; short tributaries entering the meadow from the east carried no glacial debris. The overall topographic asymmetry is probably due to faulting, but the alluvial fill which underlies the meadow, known from drilling to be as much as 200 feet thick, was probably impounded as a result of glaciation rather than faulting. The river turns from the broad deeply alluviated Big Meadow segment to flow northeast in a narrow youthful valley where rock is still commonly exposed in the channel. It seems likely that Bear Valley Creek formerly took a northerly course through the valley of Bearskin Creek to the Elk Creek drainage until it was diverted by Bull Lake glaciers. If so, the Big Meadow placer deposits may continue under the moraines and outwash that occupy the Bearskin valley; even though such deposits would not be minable under present conditions, they may be a resource for the future (Mackin and Schmidt, 1956)."

Periglacial Deposits

"Erosion and deposition in areas not scoured by Bull Lake or Pinedale ice resulted in extensive deposits including the most significant placers. These periglacial deposits were formed under a rigorous glacial climate, during

which accelerated slope erosion caused stream aggradation and colluvial accumulation (Bryan, 1949; Smith, 1949; Denny, 1951; Peltier, 1950). That such periglacial deposits formed adjacent to glacial areas in the Northern Rocky Mountains is well demonstrated in the Hailey area of Idaho (Schmidt, 1962, p. 65-69).

Monazite was dredged from tributary stream deposits along the east side of Long Valley southeast of Cascade during 1951-55. These and similar deposits were formed by streams entering the valley from a broad belt of low upland which was not glaciated during the late Pleistocene except for small ice tongues high in the headwaters 10-15 miles to the east. The degree of dissection and other criteria indicate that most of the deposits are of Bull Lake age. There is reason to believe that they were formed by aggradation caused by accelerated mass movement that resulted from the advent of Bull Lake periglacial climatic conditions. Before the periglacial aggradation a surface mantle had been enriched by selective weathering during a long period of milder climate." (Schmidt and Mackin, A13, A14, and A15)

Subsection 16 - Middle Fork Payette Canyon and Streamcut Lands Subsections

The only data available of Subsection 16 is the road log of Larsen and Schmidt (1958). This shows that much of the area is underlain by coarse quartz monzonite.

Subsection 17 - Long Valley Subsection

Subsection 17 includes Cascade Reservoir and Payette Lake. The Cascade Reservoir spillway empties into the North Fork of the Payette River. Long Valley, as mentioned in Subsection 15, is a fault block basin which has been filled with fluvial and glacial deposits. The vertical extent of these deposits is impressive. Schmidt and Mackin (1970) report over 7,000 feet of sediment in Long Valley (based on a seismic survey). Add 3,000 feet to this for the western border fault scarp of the valley which is upthrown Columbia River basalt to get 10,000 feet of throw on the north-south Long Valley Fault. The Long Valley Fault area, mentioned by Anderson (1934), extends from Long Valley 100 miles north to Grangeville and 60 miles south to Boise. This north-south fault and other parallel faults have been active in recent times as evidenced by earthquakes.

The north-south faults affected drainages as well as topography. The course of the North Fork of the Payette may have been altered by the Long Valley fault as described by Schmidt and Mackin (1970).

Origin of Drainage

"Gravel deposits in scattered exposures seen in reconnaissance at the base of the West Mountain scarp (mapped as fluvial deposits of Tertiary and Quaternary age opposite the mouth of Gold Fork) and high on the backslope of West Mountain (not mapped) consist dominantly of well-rounded smoky-quartz pebbles, mostly less than an inch in diameter. The associated monazite-bearing heavy-mineral suite was not derived from the basalt or metamorphic rocks of West Mountain but is substantially the same as the suite that characterizes similar monazite- and smoky-quartz-bearing gravel in tributaries that enter Long Valley from the east. This distinctive gravel could not have moved to West Mountain across the North Fork of the Payette River, which flows south in Long Valley; it must have been deposited by west-flowing streams before the North Fork was in its present position. Indeed, the occurrence of remnants of the gravel on the backslope of the West Mountain block indicates that it must have been deposited before that block was raised to its present height.

The most obvious hypothesis is that the original west-flowing drainage was defeated by uplift of West Mountain and that the North Fork of the Payette is, in effect, consequent on the floor of the Long Valley fault trough. Part of the river's course is, however, not entirely in accord with this hypothesis. At Cascade, the North Fork turns eastward through the raised part of the Cascade fault block and then continues south for several miles down the depressed part of the fault block. But instead of continuing in the depressed block into Round Valley, the river abruptly turns west into the southern extension of the West Mountain block and then south into a deep narrow gorge. While anomalous, this erratic behavior does not count strongly against the hypothesis; the stream could be consequent in its overall southerly course in the fault zone and locally be antecedent or superposed, or both, on minor blocks within the zone.

There is, however, an alternative possibility: the original west-flowing streams may have been blocked by the Columbia River Basalt, and the North Fork may be consequent in the topographic low between the regional westward slope on the crystalline rocks and the eastward initial slope of the basalt. This is the origin of the course of the Columbia River around the northern and northwestern borders of the same lava field. According to this second hypothesis, the monazite- and smoky-quartz-bearing gravel would be part of the Latah Formation.

> The two hypotheses are of course not mutually exclusive; the North Fork could have been consequent along the margin of the basalt and further deranged by postbasalt faulting. The hypotheses have different implications as to whether the exotic gravel is above or below the basalt on West Mountain and could be resolved by mapping." (Schmidt and Mackin, pg. A6, A7, and A8)

Glacial and periglacial sediments in Long Valley are similar to the descriptions given in Subsection 15. Most of the north-south basins in western Idaho are similar to Long and Bear Valleys.

Subsection 18 - Long Valley Foothills Subsection

Subsection 18 is within the area mapped by Schmidt (1964). He observed that rocks of the Idaho batholith graded from gneisses and schists at the western contact to migmatites, then quartz diorite gneiss, to quartz diorite, granodiorite, and finally quartz monzonite westwards. Three of these zones are present in Subsection 18, quartz diorite gneiss (mKi_2a) in the western part of the Subsection, quartz diorite (Ki_2a) in a northeast trending band that includes Little Valley, and granodiorite (Ki_3b) which underlies most of the eastern portion of the Subsection. This sequence of rocks has quite distinctive textural changes as can be seen in Figure 12. Schmidt described several endomorphic changes in the batholith rocks which were similar to Anderson's (1942) descriptions (see introductory batholith discussion).

The road log of Larsen and Schmidt (1958) cuts across the southern portion of the Subsection and they classified the entire area as coarse quartz monzonite (Ki_3ch). Schmidt's latter, more detailed report is probably more accurate than the road log. Schmidt noted several large schist inclusions (mpCl) in the batholith and these like the Thompson Peak Formation in the Sawtooths are probably Precambrian Belt equivalents.

Subsection 19 - Upper South Fork Salmon Subsection

Subsection 19 contains the headwaters of the South Fork of the Salmon River and Warm Lake. Schmidt (1964) included almost the entire Subsection in his quartz monzonite zone. The road log of Larsen and Schmidt (1958) also classifies this portion of the batholith as quartz monzonite in composition. An examination of the linear maps for the Baker and Challis AMS sheets shows that fewer linears were noted in this area as compared to Subsections further east. The few linears which are present strike north-south and are long and consistent. These linears are probably faults related to the development of the valley which contains Warm Lake.

Subsection 20 - Fitsum Peak Subsection

A section of Larsen and Schmidt's (1958) road log traversed the northern part of Subsection 20. They show a western zone of quartz diorite (border zone of the batholith) which grades to coarse quartz

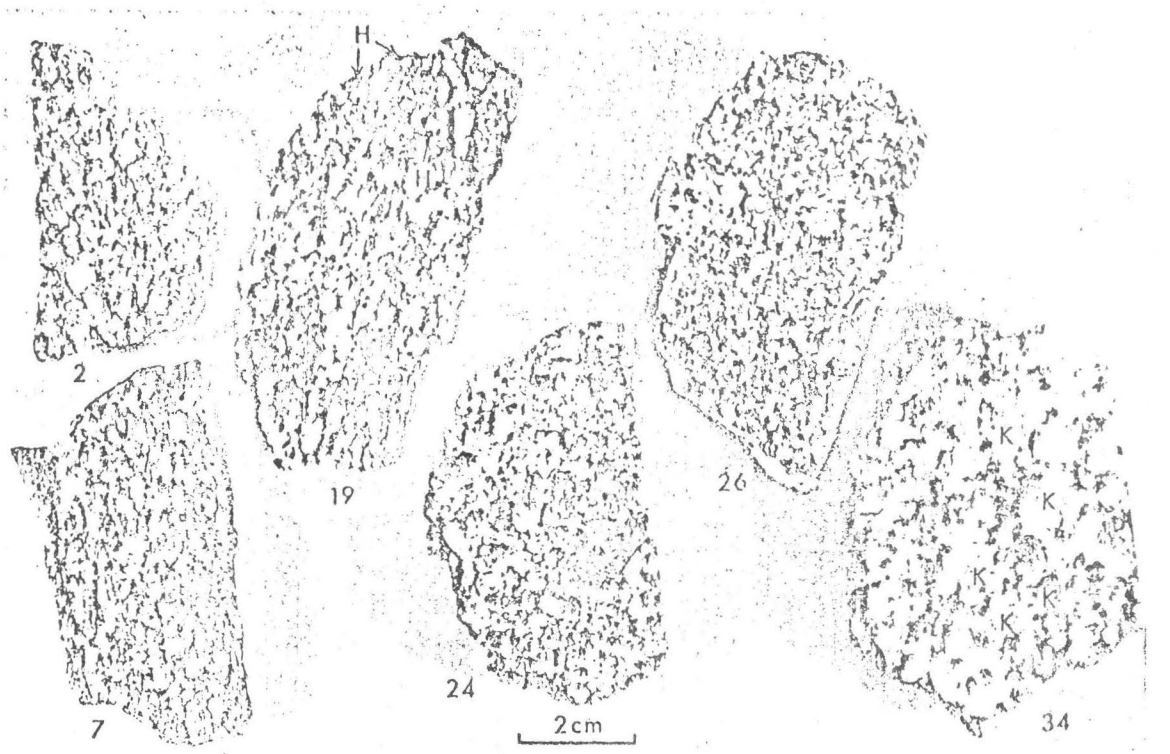


Figure 12. Polished thick slices from the gradational rock sequence across the Idaho batholith near Cascade, Idaho with legend (Schmidt, 1964).

monzonite and finally to fine quartz monzonite in the eastern part of the subsection. This is the same pattern of rock type distribution as is generally observed near the western edge of the batholith as noted in Section I, Subsection 18 and 19, south of Fitsum Peak and in Section III, Subsection 2, north of Fitsum Peak.

Subsection 21 - Upper South Fork Salmon River Canyon Land Subsection

Subsection 21 contains a major part of the South Fork of the Salmon drainage area. The road log of Larsen and Schmidt (1958) classifies the entire drainage as being underlain by coarse quartz monzonite (Ki_3ch).

Subsection 22 - Upper Middle Fork Steep Canyon and Streamcut Subsection

Subsection 22 is an area of rugged topography which is drained by the Middle Fork of the Salmon River. The main rock units in the area are Idaho batholith granitics in the western half with volcanics and Precambrian metasediments in the eastern half. A large Tertiary intrusive, the Casto pluton, occupies part of the northcentral portion of the Subsection. Ross's 1934c paper on the Casto quadrangle is the most often quoted on this area and many of the following observations are based on his work.

The Precambrian metasediments in the eastern part of the Subsection are the Hoodoo quartzite and the Yellowjacket Formation which are described in detail in the next subsection. Ross recognized another metamorphic unit near the Boyle ranch which will be referred to as the Boyle ranch sequence (mpCs). This formation consists of an upper schist unit, a section of dolomitic limestone and a lower marble which is interbedded with quartzite. All of this formation is recrystallized. The Boyle ranch rocks are similar in lithologic description to rocks in the Stibnite area which are given the same symbol. Ross described another schist unit which outcrops southeast of the Boyle ranch sequence. This unit is given the symbol (mpCl) which is the same as the Thompson Peak Formation in the Sawtooths and schist units near Cascade.

Ross described the usual rocks of the Challis volcanics including undifferentiated volcanics (Tev), the Germer tuffaceous member (Tov), and the Yankee Fork rhyolite (Tme_3). Ross further described what he thought was a sequence of Permain volcanics which he called the Casto volcanics. The major difference between the Casto and the Challis volcanics was that the Casto were highly altered and metamorphosed. Ross thought that the Casto might be the same age as the Seven Devil's volcanics in western Idaho. Cater et al (1973) have proposed that the term Casto volcanics be abandoned as the Casto appear to be nothing more than hydrothermally altered Challis. The two units grade back and forth between each other in the field. Because the Casto volcanics are so altered and are therefore distinctive they are given the map symbol Pm to distinguish them from the Challis volcanics which are unaltered.

Ross described the Idaho batholith in the Casto quad as being quartz monzonite with some small bodies of what he called white quartz monzonite which were intruded into the quartz monzonite of the batholith. The white quartz monzonite is probably equivalent to the leucocratic quartz monzonite of Reid (1963), Dover (1969), and Bennett (1973).

A large Tertiary intrusive (Casto pluton) strikes northeast from the central area of Subsection 22. This is a pink granite and is similar in description to the Sawtooth batholith pink quartz monzonite of Reid (1963). The intrusive cuts lower Challis volcanics and has a great many Tertiary dikes related to it.

A major fault system in the volcanics strikes N. 32° E. while a minor set strikes N. 27° W. and is probably a complement of the northeast set. Ross offered an explanation for the formation of Tertiary faults, intrusions, and dike swarms. The Casto pluton intruded along a series of major northeasterly faults. As the intrusive domed upwards it fractured its roof allowing the escape of magma which formed the dike swarms associated with the intrusion. Anderson (1939) noted that the roof fractures may also have been caused by the collapse of the intrusive dome. The Challis volcanics have been folded into a northeasterly trending anticline along a broad arch which extends beyond the Casto quadrangle. Whether the intrusion produced the folding or whether the folding was caused by tectonic forces on a regional scale is not known.

Ross described the Old Tertiary Erosion Surface (Oligocene?), which formed over much of the entire state, as being relatively flat in this Subsection whereas southeast of here the erosion surface had considerable relief. After the Challis volcanic episode, Ross believed erosion produced a mature relief in the area but before peneplanation occurred, regional uplift began and this has continued to the present. Ross estimated that 3,000 to 5,000 ft. of uplift had occurred since the Pliocene (based on the differences in present river elevations versus the Tertiary erosion surface). Ross recognized several stages of glaciation which had affected the Subsection.

Treves and Melear (1953) reported on the Seafoam District which adjoins the southeast corner of the Casto quadrangle. The entire district is underlain by Idaho batholith with a few small outcrops of the Boyle ranch sequence near Sheep Mountain. They recorded two major joint sets in the batholith which trended northwesterly and northeasterly. Several small tertiary dike swarms were also mentioned.

Subsection 23 - Profile-Monumental Subsection

Subsection 23 is located in the north central area of the Challis AMS sheet. It is drained by the tributaries of the Middle Fork of the Salmon in the south, by Big Creek in the north, and by the South Fork of the Salmon in its central region. Extensive areas of the central and southern portions of the Subsection are covered by Challis volcanics. The northern third of 23 is covered with Precambrian metasediments and several Precambrian intrusive bodies. There is a

radical change in topography between the northern part of the Subsection and Section III, Subsection 3. This is due to the quartzites in the Precambriam metasediments which are relatively resistant to erosion in comparison with the granitic rocks to the north.

The Yellowjacket Formation (MpCY) and the Hoodoo quartzite (mpCHo) as well as the Precambrian intrusive complexes (mpCil) were described by Cater et al (1973) as follows.

Yellowjacket Formation

"The Yellowjacket Formation of the Belt Supergroup consists of argillite, siltite, impure quartzite, and metavolcanic rocks. The formation was named by Ross (1934, p. 16) for exposures in the Yellowjacket mining district, just east of the primitive area. Almost half of the belt of Precambrian rocks that trends east-southeastward across the primitive area, along the canyons of Big and Wilson Creeks, is assigned to the Yellowjacket. Small isolated masses of metamorphic rocks in the southern part of the primitive area also are assigned to the Yellowjacket.

The Yellowjacket in the primitive area is dominated by argillite, siltite, and quartzite that is mostly impure and fine grained. The argillite and siltite are medium- to dark-gray fine-grained aggregates of quartz, biotite, and muscovite. Bedding and other sedimentary structures are visible in many outcrops despite the complete recrystallization and the common presence of a distinct cleavage. The intercalated quartzites are lighter colored, coarser, variably white to medium gray, and, also, are composed mostly of quartz, biotite, and muscovite, although they have markedly more quartz than do the argillite and siltite. Calcareous interbeds in the formation are abundant only on the ridge between Little Marble Creek and the settlement of Big Creek at the West edge of the area. There the beds consist of calcareous quartzite and quartzose marble containing calcite, quartz, and variable but generally small amounts of phlogopite, biotite, epidote, chlorite, actinolite, and plagioclase.

The Yellowjacket has been most closely studied in the Big Creek and Yellow Pine quadrangles, which include the west edge of the southern two-thirds of the primitive area, by Leonard (1962; unpub. data). Despite obscuration by faulting, isoclinal folding, and flowage accompanying metamorphism, Leonard regards the Yellowjacket there as divisible into three units. The lower unit, at least 5,000 feet thick, consists of thin- to thick-bedded argillite and siltite with scattered interlayers of fine- to medium-grained argillaceous quartzite. The middle unit, about 2,500 feet thick, consists of inter-

bedded sandy carbonate and quartzite. The upper unit, about 500 feet thick, consists of three subunits of metavolcanic rocks. In sequence upward, these subunits are (1) porphyroblastic hornblende schist, (2) interlensed hornblende and biotite schists (derived from lavas?) and hornfels (from volcanic siltstone?), and (3) metamorphosed volcanic breccia and conglomerate.

Just east of the primitive area, Ross (1934, p. 16) measured a section in rocks, metamorphosed only slightly, which displayed ripple marks and crossbeds, indicating the stratigraphic top. He found, there, a lower, calcareous unit, of which he saw about 1,700 feet, and an upper unit of argillaceous quartzite, about 7,000 feet thick, overlain in turn by Hoodoo Quartzite.

Nothing is known about the internal stratigraphy of the Yellowjacket in the Precambrian belt along Big and Wilson Creeks, which connects the areas studied by Leonard and Ross. The Yellowjacket of the belt consists mostly of argillite, siltite, and impure quartzite, and corresponds lithologically to Leonard's lower unit and to Ross' upper one. Leonard's calcareous middle unit might correspond to Ross' lower unit. Either the Yellowjacket lacks a consistent stratigraphy through the region, or that stratigraphy has yet to be defined, or the rocks designated as Yellowjacket represent similar lithologies in a thick stratigraphic section consisting of alternating quartzites (Hoodoo) and finer grained sediments (Yellowjacket).

The base of the Yellowjacket has not been recognized in the primitive area or elsewhere. B. F. Leonard (unpub. data), in the west, and Ross (1934, p. 16) in the east, both regard the Yellowjacket as being overlain stratigraphically by the Hoodoo, but, as just noted, this order need not represent a single succession."

Hoodoo Quartzite

"Masses of quartzite assigned to the Hoodoo Quartzite of the Belt Supergroup are interspersed with masses of finer grained rocks assigned to the Yellowjacket throughout the Big Creek-Wilson Creek Precambrian belt. The Hoodoo was named by Ross (1934, p. 18), for exposures along Hoodoo Creek just east of the primitive area. The Hoodoo is dominated by massive fine- to medium-grained vitreous quartzite that is white or nearly white. Bedding and crossbedding are seldom seen, having been obliterated by metamorphism in most of the formation. Interlocking quartz grains form most of the rock, but grains of plagioclase, potash feldspar, sericite or muscovite, biotite, and chlorite may total as much as 25 percent of the rock.

The quartzite is resistant to erosion and, so, tends to form ridges and knobs. But it is closely jointed, and surfaces underlain by it are mostly covered by angular fragments. Quartzite containing faintly visible pebbles, also of white quartzite, was seen in numerous places. Dark phyllite and siltite form interlayers in parts of the formation."

Age and Stratigraphic Relationship of Yellowjacket and Hoodoo Formations

"Both the Yellowjacket Formation and the Hoodoo Quartzite are intruded in the Big Creek-Wilson Creek belt of Precambrian rocks by syenite and other igneous rocks, discussed subsequently and known to be of late Precambrian age. Both Yellowjacket and Hoodoo are thus also Precambrian.

Near the west edge of the primitive area, Leonard (unpub. data) studied in detail the Yellowjacket and the Hoodoo, which lies above it in a syncline. He concluded that the relationship of cleavage, minor folds, large folds, and bedding there indicates that the Hoodoo lies stratigraphically, as well as structurally, above the Yellowjacket. He further inferred that the contact is probably an unconformity, although structural complications could, instead, have caused the lithologic and structural discordances observed. Ross (1934, p. 16) studied a section just east of the primitive area where the direction of younger beds is indicated by crossbeds and ripple marks and found that the Hoodoo there apparently also overlies the Yellowjacket. The Yellowjacket and Hoodoo each represent a single formation, and, apparently, the Hoodoo is the younger. Neither the base nor the top of the combined Yellowjacket-Hoodoo section has yet been recognized.

The Yellowjacket and Hoodoo occur in large and small masses interspersed throughout the Big Creek-Wilson Creek belt (pl. 1). Our reconnaissance mapping suggests that the contrasted types, in places, are interfolded isoclinally about steep-dipping planes but provides no basis for identifying synclines, anticlines, or stratigraphic successions. Neither the Yellowjacket nor the Hoodoo represents a particularly distinctive lithology, and, as noted previously, the little information available on the internal stratigraphy of the Yellowjacket at opposite ends of the Precambrian belt leads to contradictory conclusions. The possibility should thus be kept in mind that the Yellowjacket and the Hoodoo may represent not single formations, but contrasted lithologies in a series of formations of the two types in alternation.

The Yellowjacket and Hoodoo have long been assumed (as, by Ross, 1934) to be part of the Belt Supergroup, which is widespread over western Montana and northern Idaho. Specific correlations with other formations in the Belt Supergroup cannot yet be made, but according to E. T. Ruppel (oral commun., 1971), the Yellowjacket is similar to, and probably correlates with, the lower part of a very thick sequence of Precambrian sedimentary rocks extensively exposed in the northern part of Lemhi Range, 40 or 50 miles east of the primitive area. Obradovich and Peterman (1968) found the Belt in western Montana to span the age interval from about 900 to 1,300 m.y. ago; so, presumably, the Yellowjacket and Hoodoo would fall within this age span."

Intrusive Complexes

"The youngest known Precambrian rocks of the primitive area are complex intrusions into the Yellowjacket and Hoodoo Formations of the Big Creek-Wilson Creek belt. Each of the four intrusive masses within the belt is dominated by syenite and quartz syenite and contains subordinate hornblendite and syenite contaminated by very coarse grained amphibolite. All but the westernmost mass contain also much gabbro and some pyroxenite. The various rock types occur in part as discrete masses of considerable size and in part as intricate small-scale mixtures of varied types. The westernmost intrusive mass was called the Ramey Ridge Complex by Leonard (1963).

The quartz syenite and syenite generally are coarse grained and differ only in the presence or absence of quartz. Microantiperthite is by far the most abundant feldspar, but microperthite and nonperthitic microcline occur locally. Biotite is generally more abundant than hornblende. Magnetite, apatite, zircon, titanite, fluorite, and allanite are accessory. Chevkinite, a complex Ce-Ti silicate, was recognized by Leonard in the westernmost mass. Syenites contaminated by amphibolite are characterized by large but varying amounts of hornblende, and such rocks grade into nearly pure hornblendite.

Medium-grained gabbro is common in the three eastern masses and consists mostly of labradorite, clinopyroxene, and uralitic hornblende. Magnetite, biotite, and titanite are accessory, and actinolite and epidote are secondary. Pyrite is present but sparse.

The pyroxenite is nearly black and coarse grained and consists largely of clinopyroxene that has been partly

replaced by uralitic hornblende and fibrous actinolite. Apatite is abundant, magnetite is less so.

Discordant lead-uranium and lead-thorium age determinations for rocks of the westernmost complex, the Ramey Ridge, indicate a minimum age of 680 m.y. (Leonard and Stern, 1966). The most likely age is about 725 m.y. The syenites of the other three complexes resemble those of the dated one, and, presumably, all are of similar age." (Cater, p. 14, 15, 16, 17, and 18)

The Thunder Mountain mining district (about 1 inch below the top of the Challis AMS sheet at 115° 10' longitude) is described by Leonard (1973) and is of interest because Thunder Mountain is a caldera which was a source for the Challis volcanics. Two types of Idaho batholith rocks were described, quartz monzonite which Leonard said graded between granodiorite and quartz monzonite and mixed granitic rocks (Ki_3) which were described as being primarily alaskites with some Precambrian metasediments and metavolcanics mixed in.

The general structural trends in Subsection 23 are described by Leonard et al (1973) as follows.

Structure of Cretaceous Rocks

"Few faults were recognized in the plutonic Cretaceous rocks of the primitive area. The presence of numerous faults within the younger Challis Volcanics suggests, however, that minor faults are common within the Mesozoic rocks also, where they went unrecognized because of the lack of distinctive rock units and good exposures.

The large Bargamin Creek fault, along the northwest edge of the area, produces a lineament that is conspicuous on aerial photographs and that controls the courses of several creeks. The displacement on the fault must be large, for it juxtaposes large masses of dissimilar crystalline rocks. The large displacement and the straightness of the fault through the primitive area, and beyond it to the northeast in the Salmon River Breaks Primitive Area mapped by P. L. Weis (oral commun., 1967), indicate that it may be a strike-slip fault."

Structure of Tertiary Rocks

"The Challis Volcanics display gentle to moderate dips resulting from deformation by open folds and normal faults. The faults recognized in the field, which are conspicuous also as lineaments on topographic maps, have a dominantly northeastward trend.

The dikes in the Tertiary dike swarms also have dominantly northeastward to north-northeastward trends, whatever the rocks into which they were injected. The region apparently was undergoing northwestward extension during the early Tertiary time of dike injection. Much of the northeast-trending normal faulting of the Challis Volcanics presumably reflects this same extension."

Uplift and Erosion

"The primitive area had an altitude of about 4,000 feet during Eocene time, according to Axelrod's (1968) paleoclimatic analysis. The present crestal altitudes of about 8,000 feet thus require uplift of about 4,000 feet during middle and late Cenozoic time. Canyon cutting and, hence, perhaps late uplift in the west-central part of the area postdates the eruption of the Oligocene latite.

Part of the relative uplift of the region must be attributed to erosion itself. Incision of the canyons, which has steadily lowered the average altitude relative to the crestal altitude, must have been accompanied by compensatory isostatic rise of the region. The great canyon of the Salmon River, within the primitive area as across the rest of central Idaho, has a rim, its crest 5-8 miles from the river and 1,000 feet or so higher than the country farther back from the river. The river appears to be flowing along the crest of an anticline. Hamilton (1962) attributed this to the isostatic rise of a belt, centered on the river and at least 20 miles wide, owing to the erosion of the canyon."

Structure of Precambrian Rocks

"The older rocks of the primitive area display the most complex structure. Foliation and compositional layering are parallel in the gneiss and schist of the mass along the lower reaches of the Middle Fork. Conspicuous rodding and mineral lineations are widespread. Isoclinal-fold axes are only locally obvious, but presumably the entire mass is intensely isoclinal internally. Small-scale crumpling, crinkling, and small folding are conspicuous in many outcrops. Ptygmatic contortion of early veins of granitic material is common.

Bedding and either cleavage of foliation are not commonly both visible in outcrops of the Yellowjacket Formation or the Hoodoo Quartzite, but where both are seen they tend to be parallel or to intersect at low angles. We infer that the formations are generally tightly and in places perhaps isoclinally folded, even though very few fold

> axes were seen. The irregular variations in dips in parts of the formations are attributed to faulting and more open folding superimposed on these older structures. B. F. Leonard (oral commun., 1969) believes that in the Big Creek and Yellow Pine quadrangles, which he has mapped in detail, an early set of northwest-trending folds has been crossfolded by northeast-trending structures." (Leanard et al 1973)

Shenon and Ross (1936) noted the importance of landslide deposits in shaping the topography near Thunder Mountain. These slides often dammed up valleys which then filled with detritus. Erosion of these older fluvial deposits has produced pinnacle like remnants, one of which gave a local natural oddity, Monumental Rock, its name. Ross noted that the metasediments in the Thunder Mountain area have been deformed into northwest trending folds. Of major interest, although on a small scale is another Tertiary dike swarm, the principle zone of the swarm extends from near Profile Creek north to Sheepeater Mountain. Almost every kind of porphyry dike is present in the swarm. The belt has the same trend as a silicified shear zone which was mapped by Leonard, Mead, and Conklin (1968) which is just south of Mosquito Ridge. Several structural profiles across the Edwardsburg-Thunder Mountain area are shown in Figure 13, which shows more detail of the geology in the region.

Cooper (1950) described the mining area around Stibnite, Idaho (southwest part of Subsection 23). The major mines and prospects are localized along a silicified shear zone which also contains many Tertiary porphyry dikes. A fair sized outcrop of metasediments was described by Cooper as meta-limestones and dolomites, interbedded with mica schists. These metasediments are given the same map symbol as the Boyle ranch sequence (mpCs) because of similar lithologic descriptions.

Leonard et al (1973) mentioned a silicified shear zone north of Stibnite which trends northwest but few other particulars were given. The southeastern part of the Subsection contains large masses of the Casto pluton and the Casto volcanics which were examined in Subsection 22.

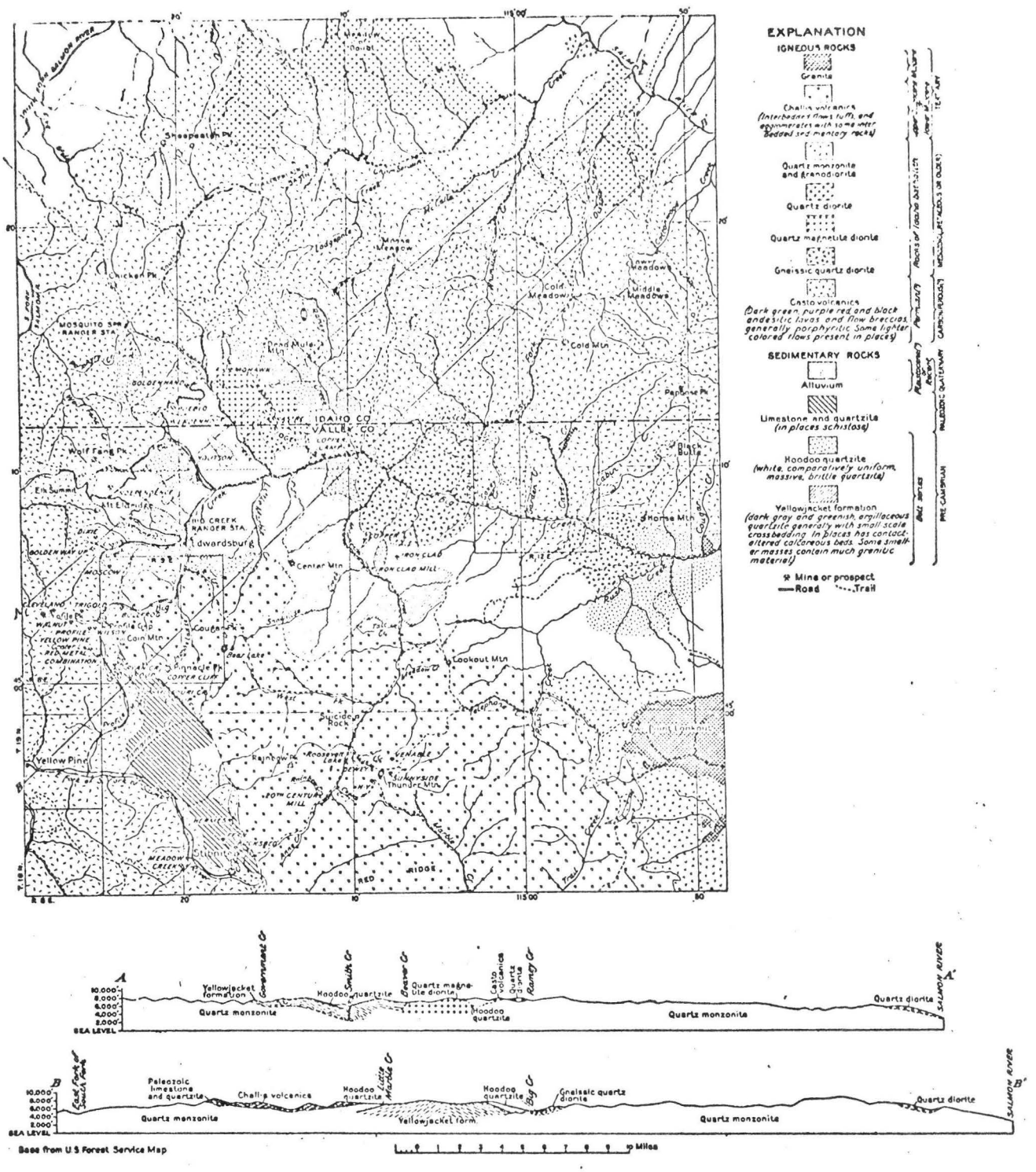


Figure 13. Reconnaissance geologic map of part of the Edwardsburg-Thunder Mountain Area (Shenon and Ross, 1936).

THE GEOLOGY OF SECTION II

CHALLIS SECTION

Section II, the Challis Section, is divided into seven Subsections. The Section is located in the eastern part of the Hailey and Challis AMS maps. A small piece of the Section extends onto the Twin Falls map but this is not included in this report. The Section is roughly included in the area of 113° 55' - 114° 45' longitude and 43° 15' - 45° 0' latitude. Major drainages within the area are the Wood River and Little Wood River in the south and the Salmon River in the north. General rock types within the Section include an extensive Paleozoic section, large masses of Tertiary volcanics, some Precambrian rocks and the Idaho batholith complex. Because the Paleozoic section and the Challis volcanics are so dominant in the Section they will be discussed now in some detail.

Paleozoic Stratigraphy

The Paleozoic Stratigraphy which is used in this report is included with the geologic map folio. All unit map symbols are shown on the stratigraphic column. The stratigraphy used is based heavily on Ross's (1962b) report. Ross first described much of the Paleozoic section in the Bayhorse area in 1937. His original descriptions have not been changed for many of these units. For this reason many of the unit descriptions are quoted here verbatim from Ross's 1962 paper.

PALEOZOIC STRATA

Cambrian System

Garden Creek Phyllite

"The Garden Creek Phyllite and overlying Bayhorse Dolomite, both of which are recognized only west of Challis, Custer County, (Ross 1937 (1938), p. 12-14) are the sole formations in south-central Idaho now tentatively regarded as of Cambrian age and even these have yielded no diagnostic fossils. This is in marked contrast to the thick and diversified marine Cambrian sequence in southeastern Idaho.

The Garden Creek Phyllite has been recognized only along the inner gorges of Garden and Bayhorse creeks in the northern part of the Bayhorse quadrangle. It is composed exclusively of a soft, easily weathered, nearly black phyllite, locally slightly calcareous, with abundant silvery sericite on cleavage surfaces. Bedding is represented by thin crenulated bands, visible on close inspection. At least several hundred feet of beds are present but marked contortion and poor exposures prevent accurate measurement of thickness.

The Garden Creek Phyllite has yielded no fossils and is provisionally referred to the Cambrian mainly because of its position at the base of the local section, well below the Ramshorn Slate (Lower Ordovician). It somewhat resembles and may conceivably be equivalent to Precambrian (?) schist in the Loon Creek district, but is distinctly more argillaceous than most of the strata in the surrounding region that are commonly regarded as belonging to the Belt Series."

Bayhorse Dolomite

"The Bayhorse Dolomite crops out at intervals along the crest and east flank of the anticline which extends diagonally across the northwestern part of the Bayhorse quadrangle. It is best exposed near the town of Bayhorse.

The greater part of the formation is composed of thick-bedded dolomite. Most of the beds are light creamy gray when fresh, but they weather readily to a rusty buff. Many are crowded with nearly black, ovoid chert masses, commonly less than half a centimeter in maximum diameter. Exceptionally, the dolomite is nearly black and, instead of the chert masses, is studded with slightly larger, white bodies. Each of these consists of a single crystalline grain of dolomite, commonly with a rim of fine-grained carbonate, at least in part calcite. Some calcite is also present in the dolomite groundmass in many of the beds of the formation. Locally there are beds of chert, quartzite, and conglomerate or breccia and some lenses, especially near the top of the formation, are argillaceous.

The formation was extensively eroded prior to the deposition of the overlying beds. Locally along Daugherty Creek it was entirely removed. The greatest exposed thickness is along Bayhorse Creek where the overlying conglomerate is absent. In the cliffs immediately north of the town of Bayhorse approximately 1,000 feet of the Bayhorse Dolomite is exposed, with Garden Creek Phyllite below and Ramshorn Slate above. On the east flank of the anticline near Garden Creek the thickness cannot be much less than 1,000 feet although here the overlying conglomerate indicates that the dolomite has been eroded.

The Bayhorse Dolomite is separated from the Lower Ordovician slate above it by an erosional unconformity which, locally at least, is of considerable magnitude. Lithologically the dolomite more nearly resembles early Paleozoic strata in neighboring regions than any known member of the Belt Series. For these reasons the formation is tentatively regarded as of Cambrian age."

Ordovician System

Summary Statement

"The Ordovician sequence is especially thick in the upper part of the drainage basin of the Big Wood River and from there northward past Challis. In the southern part of this area the system is represented by the Phi Kappa Formation, underlain in one locality by a few hundred feet of beds of earlier Ordovician age. The Phi Kappa Formation, over 9,400 feet thick, consists of dark sandy and argillaceous beds. Farther north the Ramshorn Slate, Kinnikinic Quartzite, and Saturday Mountain Formation (dominantly dolomitic) have an aggregate thickness of about 8,500 feet. In the Lost River Mountains and Lemhi Range only the Kinnikinic Quartzite and Saturday Mountain Formation are recognized. Both are markedly thinner than they are near and west of Challis; the Saturday Mountain Formation in both ranges thins southeastward. The Beaverhead Mountains contain scattered exposures of the Kinnikinic Quartzite but apparently no other rocks of Ordovician age. This statement applies also to contiguous parts of Montana."

Phi Kappa Formation

"The Phi Kappa Formation as at present defined includes all the Ordovician strata in the Hailey quadrangle with the exception of certain older unnamed beds of carbonaceous argillite, which have been recognized only in a single exposure within that quadrangle. The formation is named for exposures along Phi Kappa Creek, which enters the Big Lost River from the south, 3.5 miles down the valley from the divide, and occupies a belt that stretches southeastward from the North Fork of the Big Lost River for 12 miles and has an average width of 2 to 3 miles, mainly along the crest of the Pioneer Range. The Phi Kappa beds are in most places separated from other Paleozoic rocks by faults and are themselves cut by numerous faults, forming a complex structural mosaic. They have been disturbed and locally metamorphosed by intrusion of monzonitic stocks. In consequence of these factors and of the paucity of recognizable fossils in large areas of the rocks, subdivision of the formation is difficult. The fact that the formation is overlapped by the Challis Volcanics at several places, notably on the north, adds to the confusion and prevents direct comparison with the Ordovician strata of the Bayhorse quadrangle.

The Ordovician rocks mapped as of pre-Phi Kappa age constitute a narrow block of carbonaceous argillite along the highway down Trail Creek and about 1.5 miles

south of the pass at the head of that stream. This block is bounded by faults. It contains a few hundred feet of beds, without exposed top or bottom.

The Phi Kappa Formation may be divided for convenience of description into an upper and a lower portion, probably nearly equal in thickness. The separation of these two portions as distinct formations would have been desirable but an attempt to do so in the field proved impracticable. (ed. note - Those two units have been shown by Churkin (1962) to be in thrust contact.) The lower portion of the Phi Kappa is at least 4,750 feet thick on the west side of upper Big Lost River, where it is particularly well exposed. Somewhat over a third of this is composed of black, locally gray and white, fine-grained argillite, rather thin bedded in the lower portion, massive near the middle, and with intercalated beds of quartzitic sandstone in the upper portion. Next above is black to gray quartzite in beds 1-5 feet thick alternating with black flinty argillite. This is succeeded upward by several hundred feet of flinty, in part aphanitic, black argillite, locally spotted with incipient clusters of metamorphic minerals. Above this is hornfels, followed by more of the argillite and finally a great thickness of thin-layered, steel-gray and black hornfels. All the beds above described contain small amounts of such metamorphic minerals as tourmaline, hornblende, and, locally, pyroxene. The quartzite is crowded with rounded quartz grains 0.5-1.5 mm in diameter in a recrystallized matrix of quartz, colored black by iron oxide dust. The rock described as argillite is similar to the quartzite but finer grained.

The principal exposures of the upper portion of the formation are in the north half of the belt occupied by it, especially in the mountains east of upper Trail Creek. This portion is characterized by alternations of black laminated argillite or shale containing graptolites, and yellow shaly sandstone, in part calcareous."

Ramshorn Slate

Distribution. "All of the Ordovician argillaceous rocks below the Kinnikinic Quartzite in the Bayhorse region are grouped as the Ramshorn Slate. The formation constitutes the country rock of, and is named for, the well known Ramshorn mine. It extends in a band 2-5 miles in width from Mill Creek on the northern border of the Bayhorse quadrangle southward past Clayton, a distance of about 20 miles. More restricted and isolated outcrops of it emerge at intervals through the cover of Challis Volcanics all the way to the southern boundary of the Bayhorse quadrangle. The formation has not been recognized in any other part of south-central Idaho. Its relation to

the Phi Kappa Formation and the unnamed strata beneath that unit in the Hailey quadrangle are indeterminate on the basis of present knowledge. The most southerly exposure of the Ramshorn Slate is about 5 miles from the most northerly mapped exposure of the Phi Kappa Formation."

Character. "The main mass of the formation is composed almost exclusively of a thinbanded argillaceous rock, with well developed slaty cleavage cutting the banding at high angles. Locally, especially near Daugherty Creek, there is conglomerate at the base of the formation and some of the beds are more quartzitic or more calcareous than the average but these are everywhere in the minority. In the isolated exposures to the south, the proportion of quartzitic and carbonaceous material is somewhat greater and the slaty cleavage less perfectly developed. For purposes of description the formation may be divided into (1) the basal conglomerate, (2) the slate which is the major component, (3) the argillite and quartzite in the more southerly outcrops.

The basal conglomerate is exposed only in the vicinity of Daugherty and Garden Creeks and in a small area on upper Rattlesnake Creek. In several places small lenses of slate and of slaty quartzite interbed with and underlie the conglomerate. Most of the conglomerate is composed almost exclusively of fairly well rounded pebbles of vein quartz and quartzite with maximum diameters rarely in excess of 3 inches, in a siliceous matrix. Some of the beds are reticulated with quartz veins. The maximum thickness along Garden Creek somewhat exceeds 500 feet but in most places it is materially less.

The slate that composes the major component of the formation is a dark green to purplish, commonly conspicuously laminated rock, with local phases that are lightened in color by the presence of calcareous material or darkened by carbonaceous matter. Because the colors fade on exposure, gray tones predominate in many outcrops. Nearly everywhere slaty cleavage is well developed, so that it is rarely possible to split the rock parallel to the bedding. The quartz grains in the slate are rarely more than a few hundredths of a millimeter in diameter. Micaceous minerals, including chlorite, serpentine and biotite, are plentiful. The slate has been metamorphosed near a small granodiorite stock. Locally, lenses of limestone and quartzite are intercalated in the slate, expecially high in the sequence. Some of the quartzite and limestone beds resemble some in the overlying Kinnikinic Quartzite. The thickness of the formation in the northwest part of the Bayhorse Quadrangle is estimated to be somewhat over 2,000 feet. It may be as

much as 3,000 feet (Anderson 1954, p. 4).

In the isolated outcrops in the southern part of the Bayhorse quadrangle the rock is prevalently gray to black, weathering brownish, but some beds have the purplish cast characteristic of much of the slate farther north. In most exposures the thin banding so common farther north persists. Some of the beds are distinctly carbonaceous. These are soft, crumpled, and, unlike almost all other members of the formation, tend to part along the bedding, a fact which has permitted the discovery of the only fossils so far found in the Ramshorn Slate. Many of the beds are somewhat quartzitic and less commonly somewhat calcareous. In most of the exposures of the Ramshorn Slate in the southern part of the Bayhorse quadrangle the stratigraphically higher beds tend to include a relatively large proportion of quartzite. Some lenses in the area south of Potaman Peak resemble the Kinnikinic Quartzite, but may record a merging of the Ramshorn and Kinnikinic formations. Both here and on East Pass Creek near the southern border of the Bayhorse quadrangle the more quartzitic beds are locally shattered by tectonic forces into breccia composed of sharply angular fragments commonly less than an inch in maximum dimension."

Kinnikinic Quartzite

Distribution. "The name Kinnikinic Quartzite (Ross 1937 (1938), p. 17-18) is taken from the creek of that name that enters the Salmon River at Clayton in the northwestern part of the Bayhorse quadrangle. The formation is exposed at intervals from the vicinity of that stream northeastward to Round Valley and thence down the valley of the Salmon River beyond the mouth of Morgan Creek almost to Ellis. The various exposures northeast of Round Valley were not formally named in previous reports but their characteristics and relations are such that they clearly belong to the Kinnikinic Quartzite. That part of the Lost River Range within the Borah Peak quadrangle has small exposures of the formation on both flanks. The Donkey Hills in the same quadrangle are crossed by exposures of the Kinnikinic Quartzite.

In the part of the Lost River Range south of the Borah Peak quadrangle the Kinnikinic Quartzite has not been mapped. It is undoubtedly present at and near the mouth of Elbow Canyon near Mackay and on the second ridge east of Arco (Ross 1934b, p. 948-949; Umpleby 1917, p. 24); presumably also in other parts of the range not yet studied.

In the Lemhi Range the formation is exposed along the southwestern flank with some interruptions, from a

little north of latitude 44° 20' to near latitude 43° 53'. Near the north end of this series of exposures the formation crosses the range to the vicinity of Gilmore. It has not been found so far along the northeastern slope of the range much south of Gilmore. North of that town there are various scattered exposures in the Lemhi Range. One of these is in Ts 18-19 N, Rs 20-21 E, on the slopes above the Salmon River. Anderson (1961, p. 25-26) reports several exposures in the western part of the Lemhi quadrangle. The third one of these, which is in the valley of the Salmon River, is the largest exposure recognized north of the general vicinity of Gilmore.

The Beaverhead Mountains contain a number of scattered outcrops of quartzite that seem clearly to be correlatable with the Kinnikinic Quartzite. Rock of this character has been mapped near Nicholia (Anderson, Wagner 1944, p. 6-7) and has been noted at intervals from that vicinity to the southern end of the range (Scholten 1957, p. 159)."

Character. "In the Bayhorse quadrangle (Ross 1937 (1938), p. 17-18) most of the Kinnikinic Quartzite is a well bedded, nearly pure quartzite in which shaly partings and subordinate amounts of somewhat shaly beds are common. Most of the latter are somewhat calcareous. The prevalent color of the fresh quartzite is nearly white, but in most exposures some or all of the beds have a distinctive lavender cast that is useful in recognizing isolated outcrops of the formation. Many exposures are stained brick red, largely through the oxidation of flakes of specularite which are commonly sparsely disseminated through the quartzite and in quartz veinlets. The rock consists essentially of detrital quartz grains 0.1-0.5 mm in diameter, slightly enlarged by deposition of secondary silica, in a siliceous cement containing sericite.

In a number of places in the Bayhorse quadrangle, mainly in the vicinity of Clayton, the formation contains irregularly lenticular aggregates of impure magnesian limestone and argillaceous beds that are large enough to be distinguished in mapping. Conglomerate was noted in the formation in this region only in the vicinity of Little Bayhorse Creek.

The thickness of the Kinnikinic Quartzite in the Bayhorse quadrangle has not been directly measured. It is estimated to be about 3,500 feet thick, possibly more in places.

In the Borah Peak quadrangle (Ross 1947, p. 1102-1104) the formation is similar to and in part coextensive with the rocks just described. Most of it consists of rather

thick-bedded to massive, nearly white, pure quartzite. This consists almost exclusively of quartz, mainly in detrital grains which range up to about half a millimeter in diameter. In some of the rocks the grains are well rounded but in others they are irregular. Quartz added in optical continuity with the original grains is locally present but conspicuously less so than in the underlying Swauger Quartzite. Some of the rocks show closely spaced lines of fluid inclusions in the detrital grains. Almost the only constituents other than quartz are the sericite flakes in a few of the quartz grains and rare grains of brown tourmaline. Some of the quartzite is stained with iron oxides. At the top of the formation joints along bedding are relatively closely spaced and locally the quartzite is interbedded with dolomite of the overlying formation, some of which is partly silicified. In a few places near the base there are narrow beds of rusty siliceous dolomite and of conglomerate.

In the area near the head of the West Fork of the Pahsimeroi River most of the Kinnikinic Quartzite consists of white quartzite like that described above. Some has distinct lavender tints and a few beds are as deeply colored as the lavender quartzite of the Swauger Quartzite. In parts of this area dark dolomitic and argillaceous beds appear to be included in the formation. Here the structure is so complex that stratigraphic details are difficult to determine.

In the Borah Peak quadrangle satisfactory measurements of the total thickness of the Kinnikinic Quartzite could not be made. It is surely as much as 2,000 feet and locally as much as 3,000 feet.

Most of the Lost River Range southeast of the Borah Peak quadrangle has received little geologic study. Near the mouth of Elbow Canyon highly contorted quartzite underlies dolomite containing fossils of Late Ordovician age. The character and stratigraphic position of this quartzite indicate that it belongs to the Kinnikinic Quartzite. The rock is nearly white, locally with a pinkish or lavender tinge, and forms distinct beds a foot or more thick, with a uniform texture and vitreous luster. It is composed almost exclusively of quartz in interlocking grains 0.4-0.7 mm in diameter. Locally, dolomitic quartzite beds are intercalated. The rock near Arco reported by Umpleby is similar. The thickness near and south of Elbow Canyon may be about 1,000 feet.

The Kinnikinic Quartzite in the southern Lemhi Range (Ross 1961a, p. 202-203) is megascopically much like that in the Borah Peak quadrangle already described. Conglomerate lenses are present here and there, especially near the base. High in the sequence a few lenses

of dolomite closely resembling the overlying Saturday Mountain Formation occur. The component detrital grains are fairly well rounded and range up to 0.5 mm in diameter. They merge with the cement in such fashion that on superficial inspection under the microscope the rock seems to be composed of interlocking grains of quartz with the original sedimentary texture destroyed. Crushed borders around the clastic grains, so conspicuous in the underlying Swauger Quartzite, are absent.

In the part of the Lemhi Range within the Lemhi quadrangle (Anderson 1961, p. 26) the Kinnikinic Quartzite does not differ materially from that in other localities. It is extensively eroded but as much as 1,000 feet of beds remains intact.

In a few places, mostly where exposures are limited, possibility exists of confusing the Kinnikinic Quartzite with either the Hoodoo or the Swauger Quartzite. The absence of feldspar in the Kinnikinic and its presence in the two Precambrian quartzite formations aids in discrimination where stratigraphic relations are indeterminate."

Saturday Mountain Formation

Distribution. "The Saturday Mountain Formation was named for a ridge which lies west of lower Squaw Creek near the middle of the western boundary of the Bayhorse quadrangle (Ross 1937 (1938), p. 18-22). In the Bayhorse region the formation is exposed only in the general vicinity of Squaw Creek and in a narrow band in the north-central part of T 12 N, R 19 E, farther east in the Bayhorse quadrangle.

East of the Bayhorse quadrangle the distribution is similar to that of the Kinnikinic Quartzite. In the Borah Peak quadrangle (Ross 1947, p. 1102-1105) one area of outcrops lies on the lower slopes of the northeast side of Mahogany Hill in and near the southwest part of T 13 N, R 22 E. Another crosses the Donkey Hills near the southwest corner of T 11 N, R 25 E. Irregular masses crop out between Borah Peak and the West Fork of the Pahsimeroi River. The formation has been seen in Elbow Canyon (Ross 1934b, p. 953, 955) and may be present in other unmapped parts of the Lost River Range. Dolomite that probably belongs to it exists in the Arco Hills in T 4 N, R 27 E. The formation is present near Gilmore on the northeast slope of the Lemhi Range and presumably crops out extensively along the southwest side of the range, although it has not yet been mapped there between latitude 44° 20' and 44° 10'. Exposures are irregular but nearly continuous along the southwest side of the range from 44° 10' to

about 43° 53' (Ross 1961a, p. 203-209). Anderson (1961, p. 27) notes small exposures near Lemhi and on the lower slopes of the Beaverhead Mountains."

Character. "In the vicinity of Squaw Creek in the Bayhorse quadrangle the formation is composed dominantly of more or less shaly dolomite and magnesian limestone, most of which contains enough carbonaceous matter to be black or nearly so on fresh fracture. In this vicinity many of the beds are more argillaceous than is common at the same horizon in other localities. The fine-grained, carbonaceous and argillaceous beds are commonly banded and more or less fissile, locally ripple marked, and some contain rounded pebbles of quartz and quartzite about 0.1 mm in diameter, but in most of the argillaceous beds quartz grains are only about 0.2 mm in diameter. In general, the Saturday Mountain beds in this vicinity are distinguished from the overlying Milligen Formation by a greater proportion of calcareous beds and by a lower average content of carbon, but a sample from one such bed was found by Charles Milton to contain 21.3 percent graphite (Ross 1937 (1938), p. 19). The carbonate in calcareous Milligen beds is mainly calcite instead of dolomite as in much of the Saturday Mountain Formation. In most other localities, the more argillaceous beds are absent, and the formation is composed of somewhat thin-bedded, dark-gray, locally rusty, dolomitic limestone. The strata are distinguished from the Laketown Dolomite where that unit overlies them by a more rusty appearance, somewhat darker color, and greater tendency to weather into slabs. The thickness is about 3,000 feet.

In the Borah Peak quadrangle (Ross 1947, p. 1104-1105) the Saturday Mountain dolomite is more uniform than in the areas described above. Most of it is a dull, dark-bluish-gray, almost black, somewhat impure dolomite, with poorly defined bedding. Some beds contain small irregular nodules and laminae of light-colored chert, and some have been partly silicified. Locally the dolomite is comparatively light colored and in other places the dark dolomite is mottled with light patches. In a few places a narrow zone of sheared dolomitic shale separates the massive dolomite from the underlying Kinnikinic Quartzite, and in other exposures the dolomite and quartzite are interbedded along the contact in a transition zone up to 200 feet thick. The massive dolomite above the transition zone is 500 to 700 feet or more thick in different areas.

Farther southeast in the Lost River Range the Saturday Mountain Formation has been studied only in Elbow Canyon near Mackay. At this place the dolomite consists of a little more than 400 feet of nearly black

beds with 200 feet of lighter colored ones above.

By far the greater part of the formation in the southern Lemhi Range (Ross 1961a, p. 203-209) consists of dolomite essentially identical in character with the dominant rock of the same formation in the Borah Peak quadrangle. In places the bedding is more distinct than in that quadrangle, and at some horizons light and dark beds alternate. A few beds are argillaceous and some, especially near the base, are quartzitic. Lenses of quartzitic conglomerate are present locally. These various clastic beds nowhere constitute a large proportion of the whole. In places at the base quartzite and dolomite are interbedded in a transition zone, but the passage from the Kinnikinic Quartzite into the Saturday Mountain Formation is in general more abrupt than in many parts of the Borah Peak quadrangle.

Much of the quartzite in the formation has a cement of magnesian carbonate. Rare grains of microcline and plagioclase are present in some beds, whereas feldspar was not noted in the Kinnikinic Quartzite. In the dolomite that constitutes most of the formation component grains average close to 0.01 mm in diameter, with some patches and veinlets of coarser crystals, whereas the grains in the dolomite in the Borah Peak quadrangle range up to 0.05 mm in diameter.

The thickness of the Saturday Mountain Formation varies from place to place in the southern Lemhi Range. In some of the most southerly exposures it is about 250 feet thick but a short distance to the east 520 feet were measured. Northward the thickness increases until in the vicinity of Badger Creek it is between 1,000 and 1,500 feet. Near Gilmore, farther north and on the northeastern slope of the Lemhi Range, the reported thickness is 500 feet (Umpleby 1913a, p. 33)."

Age. "The dolomite in the Saturday Mountain Formation in the Bayhorse quadrangle is not richly fossiliferous. Eight collections scattered throughout the formation yielded forms that Kirk reported represent a fauna of Upper Ordovician age widespread in the surrounding region. The horizon is equivalent to the Fish Haven Dolomite of Utah and southeastern Idaho, and the upper portion of the Bighorn Dolomite of Wyoming."

Silurian System

Trail Creek Formation

"Silurian strata occur in a number of localities in the eastern part of south-central Idaho. They are mainly dolomitic but in the Wood River and Bayhorse regions

include the argillite and sandstone of the Trail Creek Formation (Umpleby, Westgate, Ross 1930, p. 23-24) which has been recognized only in a strip west of the upper course of Trail Creek, a tributary which joins the Big Wood River at Ketchum, and in a small exposure surrounded by Challis Volcanics in sec. 28, T 12 N, R 19 E. (Ross 1937 (1938), p. 22-23).

The best known exposure of the Trail Creek Formation is on the west bank of Trail Creek about a mile south of the place where the creek swings abruptly to the west. Here there are about 500 feet of siliceous argillite and quartzitic sandstone. The contact with the underlying Phi Kappa Formation is clearly shown. In Malm Gulch the beds assigned to the Trail Creek Formation are mainly brownish-gray, calcareous argillite that breaks in thin slabs parallel to the bedding. The formation consists essentially of calcite, quartz, and plagioclase, with subordinate amounts of organic matter, muscovite, and apatite, and small chert lenses. On the northeast side of Malm Gulch nearly white quartzite resembling that which is elsewhere interbedded with the Laketown Dolomite overlies the argillite.

In the basal 60 feet of the formation on Trail Creek the quartzitic sandstone layers are separated by thin seams of carbonaceous argillite, in which was found an abundant graptolite fauna. Kirk reported that the graptolites are of Silurian age, approximately equivalent to the Niagara of the eastern United States.

The exposure in T 12 N, R 19 E, in the Bayhorse region contains Monograptus, indicating approximate age equivalence to the Trail Creek Formation in its type locality, where the fauna is more diversified. However, there is a possibility that the strata in T 12 N, R 19 E may be equivalent to some of the lower beds in the Laketown Dolomite of the same region."

Laketown Dolomite

Distribution. "The principal unit of Silurian age in south-central Idaho has been termed the Laketown Dolomite (Ross 1934b, p. 957-960) because it is regarded as essentially equivalent to the well known unit of that name in Utah and southeastern Idaho. In the Bayhorse quadrangle (Ross 1937 (1938), p. 23-25) the principal exposures extend from sec. 35, T 13 N, R 19 E, to sec. 18, T 11 N, R 20 E, but there are small exposures in other parts of the quadrangle. In the northern Lost River Range (Ross 1947, p. 1105-1107) isolated exposures of the Laketown Dolomite are scattered from unsurveyed T 13 N, R 21 E, to the unsurveyed eastern part of T 9 N, R 22 E, and in the southeastern part of

T 11 N, R 24 E, and the southwestern part of T 11 N, R 25 E. The formation is also present in the southern part of the Lost River Range but has been identified only in Elbow Canyon (Ross 1934b, p. 958). Beds that appear to belong to the Laketown Dolomite have been reported from near Gilmore (Umpleby 1913a, p. 33-34) and a few are present in the southern part of the Lemhi Range. Data at hand as to the intervening 25 miles of the range are insufficient to show the extent to which the formation may be exposed. Anderson (1961, p. 27) found beds he correlates with the Laketown in the southeastern part of T 18 N, R 23 E. The formation is absent or inconspicuous in the southern Beaverhead Mountains (Scholten 1957, p. 159)."

Character. "In the Bayhorse quadrangle most of the Laketown Dolomite consists of moderately thick-bedded, bluish-gray, dolomitic limestone, much of which weathers rusty. Chert nodules are commonly present but nowhere abundant. Some of the dolomite is somewhat sandy and some shaly. At several horizons throughout the formation, beds of light slate-gray to white, slabby quartzite are intercalated.

None of the exposures in the Bayhorse quadrangle include the total thickness of the formation but most of it is probably present in the principal exposure mentioned above. Here nearly 1,000 feet of quartzite is overlain by fully 2,500 feet of dolomite (with some intercalated quartzite) followed locally by several hundred feet of quartzite. In a small exposure in the northeast corner of the Bayhorse quadrangle the Laketown is made up of a rapid alternation of quartzite and dolomite aggregating over 3,000 feet. Some large exposures, such as that which crosses the northern border of T 12 N, R 19 E, are almost exclusively dolomite. In general, the formation in this part of the region appears to be well over 3,000 feet thick, with dolomite commonly making up materially the greater part of this total. Probably the quartzite, locally abundant, is in lenticular bodies which are individually of small areal extent and irregular distribution.

The Laketown Dolomite in the Borah Peak quadrangle is broadly similar. It consists mainly of light bluish-gray dolomite, readily distinguished from the darker and less distinctly bedded dolomite of the underlying Saturday Mountain Formation. Locally, however, the Laketown Dolomite is so dark that it cannot be definitely distinguished from the lighter beds in the Saturday Mountain Formation on the basis of lithologic appearance alone. The Laketown Dolomite is a little coarser and more uneven in texture than most of the dolomite of the Saturday Mountain Formation. Some grains are almost

0.1 mm in diameter. The quartzite in the Laketown is a fine crystalline mosaic whose appearance both in the field and under the microscope suggests that it is, in part, a replacement of dolomite. This rock is easily distinguished from the quartzite in older formations. The thickness of the formation in the Borah Peak quadrangle is somewhat more than 600 feet.

At Elbow Canyon farther south in the Lost River Range the comparable strata are light dolomite 325 feet in total thickness. Near Gilmore in the Lemhi Range beds apparently equivalent to the Laketown are reported to be 200 feet thick (Umpleby 1913a, p. 33). In the southern Lemhi Range (Ross 1961a, p. 208-209) beds believed to be of Laketown age are lithologically indistinguishable from the Saturday Mountain Dolomite and may aggregate only a few tens of feet in thickness."

Devonian System

Jefferson Dolomite

Distribution. "The name Jefferson Dolomite as applied in south-central Idaho is an extension southwestward of the term Jefferson Limestone long used in Montana (Cooper and others 1942) for carbonate rocks that are largely magnesian and have a stratigraphic position so nearly akin to that of the unit in south-central Idaho that the use of the name seems clearly justified. Rocks here assigned to the Jefferson Dolomite occupy scattered, small exposures in the Bayhorse quadrangle (Ross 1937 (1938), p. 25-27), the largest of which are in T 12 N, R 20-21 E. In the Borah Peak quadrangle (Ross 1947, p. 1107-1109) it is exposed at intervals along the southwest slopes of the Lost River Range from latitude 43° 20' to the southern border of the quadrangle, and it continues a short distance beyond that border. It is present in unsurveyed parts of T 13 N, R 21-22 E, around Borah Peak, and in T 11 N, R 24-25 E.

The Jefferson Dolomite is exposed in Elbow Canyon (Ross 1934b, p. 961) and probably is essentially continuous along the southwest border of the Lost River Range from near there for fully 10 miles. It is also exposed in T 4 N, R 29 E near Arco.

The formation is present in the Lemhi Range from near Gilmore (Umpleby 1913a, p. 34) at intervals to near the southeastern end of the range (Ross 1961a, p. 209-216) but has not been distinguished in mapping. It is probably also present, but unmapped, in the southern Beaverhead Mountains (Scholten 1957, p. 159-162). Anderson (1961, p. 28-29) found rock he regards as Jefferson Dolomite in and near the southern part of T 17 N, R 23 E."

Character. "The greater part of the Jefferson Dolomite in the Bayhorse and Borah Peak quadrangles is a distinctive, easily recognized rock. It is a dark bluish-gray dolomite which in many exposures contains an abundance of a characteristic digitate form of Favosites. The color is even more nearly black than, and is also brighter than that of the principal beds of the Saturday Mountain Formation, a fact which aids materially in distinguishing representatives of the two units in isolated outcrops. The dolomite is laminated, moderately thick bedded, and commonly is irregularly textured. In much of it the grains tend to show crystal forms and to be larger than those in the dolomite beds in underlying formations. Exceptionally the dolomite contains sand grains.

Some of the beds, expecially low in the Jefferson Dolomite, are lithologically indistinguishable from the light-gray dolomite of the underlying Laketown Dolomite, except that they may be coarser grained. These beds are a little more siliceous and less carbonaceous than the more plentiful dark dolomite. Similarly some beds in the upper part of the Jefferson Dolomite resemble the overlying Grand View Dolomite.

Yellowish-buff, massive beds of calcareous sandstone or quartzite are conspicuous just above the base of the Jefferson Dolomite in the Borah Peak quadrangle, and similar rock forms lenses at higher horizons also. The component grains are principally quartz but include some microcline. They are water worn but not very smooth, and many show check marks as though they were weathered before being incorporated into the rock. A few have strain shadows. The diameters range up to half a millimeter. The cement is calcareous and in some beds is abundant.

The descriptions given above apply in general to the Jefferson Dolomite wherever exposed in south-central Idaho. In detail, the succession varies from place to place. In the Bayhorse quadrangle a thickness of 1,150 feet was measured. In the Borah Peak quadrangle observed thicknesses range from 600 to over 800 feet. In Elbow Canyon 700 feet of beds that, according to Kirk's field study should be correlated with the Jefferson, are present. In the Lemhi Range the beds of Devonian age have not been subdivided into formations. Here the aggregate thickness ranges from about 300 feet near the southern end of the range (Ross 1961a, p. 209-216) to "at least 2,000 feet" north of Gilmore (Umpleby 1913a, p. 34). Of the Devonian rocks near the southern end of the Lemhi Range about 150 feet resemble the Jefferson Dolomite of the Borah Peak quadrangle. These have yellow, calcareous, quartzitic sandstone near the base and some of the dolomite contains stromatolites.

Somewhat north of latitude 44° in that range the entire sequence of Devonian rocks is exposed and its thickness is 1,200 feet, of which possibly as much as 500 feet may be equivalent to the Jefferson.

The rock Anderson (1961, p. 28-29) correlates with the Jefferson in the Lemhi quadrangle is similar to that in other localities in south-central Idaho but is more siliceous and locally has narrow layers rich in silica. In places where dolomite has weathered out, the formation shows extensive collapse, resulting in the production of large breccia blocks of siliceous rock. The Jefferson Dolomite in the Lemhi quadrangle may be close to 1,000 feet thick."

Grand View Dolomite

Distribution. "The Grand View Dolomite was named for the canyon of that name in T 12 N, R 20 E (Ross 1934b, p. 963). It is exposed there and in neighboring parts of the Bayhorse quadrangle (Ross 1937 (1938), p. 27-29) and at intervals along the southwest side of the Lost River Range to just south of the border of the Borah Peak quadrangle, as well as in other places in that quadrangle (Ross 1947, p. 1109-1110). The formation has not been mapped elsewhere but is probably present in the southern part of the Lost River Range and in the Lemhi Range. If it is present in the Beaverhead Mountains, it has not been definitely recognized. Anderson (1961, p. 29) reports a small amount of rock he regards as Grand View Dolomite in and near sec. 1 and 2, T 17 N, R 23 E."

Character. "The Grand View Dolomite is similar to the Jefferson Dolomite but contains proportionately fewer nearly black beds and fewer beds of sandstone or quartzite. Much of it is moderately dark dolomite that weathers rusty. It is commonly in beds 1-3 feet thick and is more distinctly bedded than the Jefferson. Variations in color are more marked than in the Jefferson. Nearly black beds are more plentiful in the lower part of the Grand View than above but some occur near the top of the formation. Many of the dolomite beds are somewhat sandy and there are rare beds of quartzite and shale. In Grand View Canyon a thickness of 1,170 feet was measured; the upper contact is obscured by alluvium so the total thickness slightly exceeds that figure. In unsurveyed T 13 N, R 21 E, the thickness is still greater. The beds measured total 2,115 feet and the topmost beds are not well enough exposed to include in the measured section. Near the southern end of the Lemhi Range beds that resemble the Grand View Dolomite are only about 150 feet thick but farther north the thickness may be nearly twice this. Anderson says the thickness may exceed 500

feet in the area in which he noted the unit."

Three Forks Limestone

Distribution. "Throughout the Borah Peak quadrangle a thin formation, not recognized farther west, separates the Grand View Dolomite from the Milligen Formation. Because this formation resembles both lithologically and in age the Three Forks beds (Peale 1893, p. 29-32), it is called the Three Forks Limestone (Ross 1947, p. 1110-1112). It contains significantly less shale than the corresponding unit in Montana (Cooper and others 1942, p. 1784; Sloss, Moritz 1951, p. 2152). Within and immediately south of the Borah Peak quadrangle the Three Forks Limestone has essentially the same distribution as the two formations of Devonian age already described. Like them, it appears to be present in Elbow Canyon and presumably other localities in the southern part of the Lost River Range. Rocks of this age are certainly present in the southern part of the Lemhi Range and presumably occur in most localities throughout that range in which rocks of Devonian age are present. It has not, however, been mapped separately there. Beds of Late Devonian age are present in the Beaverhead Mountains (Shenon 1928, p. 7) but whether these belong to the Three Forks or an older unit is not known."

Character. "In and just south of the Borah Peak quadrangle (Ross 1947, p. 1110-1112) the Three Forks Limestone consists mainly of more or less argillaceous limestone with some calcareous shale and locally a little quartzite. Many of the beds, especially the more shaly ones, weather such bright yellow as to be readily recognized from a distance. This and the locally abundant fossils make the unit a useful horizon marker even where thin and poorly exposed. The details of lithology vary markedly from place to place. The thickness ranges from less than 30 to more than 300 feet. The formation may be gradational with the overlying Milligen Formation.

Although the Three Forks limestone has not been mapped in other parts of south-central Idaho it doubtless occurs. In the southern part of the Lemhi Range, especially, fossils (Ross 1961a, p. 209-216) show the beds corresponding in age to this unit are widespread. These beds are mostly argillaceous limestone with some calcareous shale. Thicknesses range from 30 to possibly as much as 250 feet and increase northward."

Carboniferous and Permian Systems

Outline of Stratigraphic Relations

"The mutual relations of the dominantly marine strata of late Paleozoic age are only beginning to be understood (Ross 1962). Most of the formations as now mapped have excessively long age ranges and the gradations and interfingerings among them offer formidable obstacles to subdivision, expecially as fossils are scarce in some critical localities. East of the Idaho batholith, the dominantly marine beds include the Milligen Formation, which is mostly of late Mississippian age; the White Knob (formerly Brazer) Limestone, thought to range in age from Early Mississippian into the Permian; and the Wood River Formation, long regarded as Pennsylvanian, but probably including beds of Permian age. The newly recognized Copper Basin Formation interfingers with and locally takes the place of the Milligen and Wood River Formations and the White Knob Limestone. It thus ranges in age from Early Mississippian into the Permian. The Phosphoria Formation (Permian) has been recognized in one locality near the border of Montana. The cumbersomely thick units, other than the Phosphoria, record fairly continuous marine deposition during late Paleozoic time in the northern end of a major geosyncline. Future work will result in many subdivisions corresponding to lithologic differences that resulted from fluctuations in conditions of deposition. Many of these subdivisions will be local only and, unless new methods of dating are perfected, difficulties in establishing regional time-stratigraphic subdivisions are to be expected. Some map units based on lithologic characteristics will be found to transgress time." (Ross's descriptions)

An excellent review of the Milligen Formation was presented by Paull, Wolbrink, Volkmann and Grover in 1972.

Distribution and Thickness. "The Milligen Formation is a widespread stratigraphic unit in south-central Idaho that was mapped and is described in the Big Wood River Valley (Umpleby et al., 1930), Bayhorse region (Ross, 1937), White Knob Mountains (Skipp and Mamet, 1970), and in the Lost River and Lemhi Ranges (Ross, 1947, 1961a). The Milligen in the central Pioneer Mountains was mapped in the Mackay 3NW (Wolbrink, 1970) and 3SW (Lukowicz, 1971) quadrangles, and in the Hailey 1NE (Harry Canyon) quadrangle (Pruit, 1971). In all these areas, it maintains a predominantly dark-gray argillaceous character, although other lithologic types are present.

Kiilsgaard (1950, p. 40) reported a minimum thickness of 5,500 feet and a maximum thickness of 7,500 feet for

the Milligen in Parker Gulch in the Big Wood River valley area, although the base was not exposed. The Milligen Formation thins east and southeast of the Pioneer Mountains in the Lost River and Lemhi Ranges (Sandberg et al., 1967).

In the central Pioneer Mountains, all measured sections of the Milligen Formation are incomplete because the stratigraphic base of the formation is not exposed. The maximum measurable section is the proposed reference section, which is 3,200 feet thick. Below the measured section there are 500 feet of poorly exposed Milligen rocks that are probably conformable with the sequence above. If so, the minimum thickness in this area is 3,700 feet. Lukowicz (1971) reported an incomplete Milligen section from 4 mi. south of the reference section that is at least 3,460 feet thick. Lesser thicknesses were measured in other areas within the central Pioneers."

Character. "The Milligen Formation is predominantly dark-gray (N3), thin- to thick-bedded argillite, with minor interbedded quartzite and granule-size conglomerate. Quartzite is present sporadically throughout, but conglomerate is present only in the middle or upper part of the formation. Four miles south of the reference section, calcareous sandstone and argillaceous, micritic limestone interbedded with argillite are present through a 400 foot interval about 2,000 feet below the top of the formation. This unit was mapped as an unnamed member of the Milligen by Lukowicz (1971).

Argillite is used in this paper as defined by Twenhofel (1937) and refers to argillaceous rocks that do not possess slaty cleavage but are more highly indurated than mudstone, siltstone, claystone, or shale. The argillite of the Milligen is dark gray (N3) to medium light gray (N6), and consists of 50-90 percent clay-size matrix of illite and some chlorite, and 10-50 percent silt-size quartz, chert, and siliceous rock fragments that commonly are concentrated as laminae. Very fine-grained sand-size particles similar in composition to the silt may make up as much as 20 percent of the argillite. Minor amounts of carbonaceous material and pyrite are also present. The rock weathers moderate brown (5YR 3/4) with a dull luster and irregular to subconchoidal fracture that produces either sharp, blocky talus or grayish, earthy slopes with chips of argillite.

The argillite generally is well bedded but, locally, jointing and blocky weathering may obscure bedding. Bedding thickness ranges from 3 in. to 5 ft., but generally is 6 in. to 1 ft.

The interbedded quartzite is dark gray (N3) to dark yellowish brown (10YR 4/2) and weathers moderate brown (5YR 5/6) to medium dark gray (N4). Most quartzite beds are 1-5 feet thick.

The quartzite generally contains silt-size to fine-grained sand-size particles, but may contain medium- to coarse-grained sand. The grains appear to be subangular to subrounded and moderately well sorted, but grain relations are partly obscured by pressure-solution effects. The quartzite studied in thin section contains 75-80 percent quartz, 5-15 percent epidote (clinozoisite?), 0-3 percent plagioclase and chert, and 5-10 percent clay-size matrix and opaque material.

The conglomerate is medium gray (N5) to dark gray (N3), and weathers dark yellowish brown (10YR 4/2). Most conglomerate beds are poorly sorted and are granule-size with a few small pebbles. Beds are 3-8 feet thick and locally graded, but the grading is difficult to discern because of the dark color of both the matrix and the clasts. The point count of one thin section indicated 33 percent medium- to coarse-grained quartz sand, 39 percent medium-grained to granule-size chert, 15 percent granule-size rock fragments of quartzitic siltstone, and 13 percent siliceous and argillaceous matrix. Many of the chert clasts contain poorly preserved radiolarians." (Milligen's description)

The Milligen Formation is overlain in the Mackay 15 minute quadrangle by the Copper Basin Group. This group used to be called the Copper Basin sequence (Ross, 1962) and before that the Muldoon Formation (Thomasson, 1959). The five members of the Copper Basin Group are described as follows (Paull, Wolbrink, Volkmann, and Grover, 1972). All five of these units are grouped under the general map symbol for the Copper Basin Group PL.

Distribution and Thickness. "Ross (1962a, p. 385) suggested that the Copper Basin Formation might extend into the Lost River Range. However, our intent is that the Copper Basin Group, as defined, be restricted to rocks exposed within the Pioneer Mountains along a northwest-trending band, about 12-18 mi. wide, that extends from the Snake River plain to northwest of the Trail Creek Pass road. Refinement of the northern outcrop limit must await more detailed work on the upper Paleozoic rocks in the Bayhorse region originally mapped by Ross (1937).

The measurable thickness of the Copper Basin Group is 17,745 (± 1,050) feet, which is the total thickness of the six formations in the type area. The five type sections and the Milligen reference section are shown as a composite section which totals 17,135 feet, because 500 feet of a partly covered interval at the base of the Milligen Formation was not plotted and a minimal thickness of 2,040 feet was used for the Brockie Lake

Conglomerate. Because the top of the youngest formation (Iron Bog Creek) and the base of the oldest formation (Milligen) are not exposed, 17,745 (±1,050) feet is a minimum thickness for the Copper Basin Group. A similar conclusion was reached by Nelson and Ross (1969b, p. 42), who estimated the thickness of the Copper Basin to be about 18,000 feet "... based on inspection of the geologic map of the quadrangle." It is our opinion that if a complete section were available, it would not exceed 21,000 feet.

Lukowicz (1971) reported an incomplete composite thickness of about 17,000 feet, 5 mi. south of the type section in T3N, R22E. All the formations comprising the Copper Basin Group are present in this area, but the base is not exposed and there is an erosional top to the sequence. Further, several units are faulted. About 15 mi. southeast of the type section, in T2N, R23E, Bollmann (1971) documented an incomplete composite thickness of approximately 6,200 feet. All formations except the Milligen and Iron Bog Creek Formations are present at this locality, although they are appreciably thinner. About 18 mi. northwest of the type section, in T7N, R20E (unsurveyed), Pruit (1971) reported an incomplete composite section about 4,000 feet thick that did not include the Iron Bog Creek Formation and contains only a few hundred feet of the Milligen Formation. Although structural complexities in this area make thickness measurements somewhat uncertain, it is clear that all the formations present are thinner than at the type area."

Character. "The lithology of the Copper Basin Group is described in detail as each of the six formations comprising the group is discussed. The general lithologic characteristics of the group, from the base upward, are: (1) 3,700+ feet of dark argillite with some interbeds of dark quartzite and granule-size conglomerate; (2) 2,620 feet of gray, micritic limestone with interbeds of argillite and sandstone; (3) 3,625 feet of dark conglomerate and interbedded quartzite; (4) 2,150 (±800) feet of dark-gray argillite with small amounts of quartzite, conglomerate, and limestone; (5) 2,150 (±250) feet of light colored conglomerate and quartzite; and (6) 1,500+ feet of dark-gray argillite and shale with a few interbeds of conglomerate and quartzite.

In summary, more than half of the Copper Basin Group at the type section is argillite with small amounts of interbedded quartzite and conglomerate (9,015+ feet). The next most abundant lithologic type is conglomerate interbedded with quartzite (5,775 ± feet). Significant amounts of limestone (2,955 feet) are also present. Although thickness and lithologic details change north and south of the type area, relative lithologic relations are reasonably consistent throughout the central Pioneer Mountains."

Age and Stratigraphic Relations. "The Copper Basin Group is an apparently continuous sequence of at least 18,000 feet of sedimentary rocks of Early(?) Mississippian to Middle(?) Pennsylvanian age. These age assignments are somewhat conjectural and are based largely on lithologic correlation with more fossiliferous Early Mississippian to Middle(?) Pennsylvanian rocks on the east, in the White Knob Mountains and the Lost River Range. These correlations and the resulting age assignments are detailed as the individual formations comprising the group are discussed.

The general age assignment is supported by Nelson and Ross (1969b, p. 32), who observed that, 'All of the diagnostic fossils so far obtained from the Copper Basin Formation are of Late Mississippian age. However, some collections contain forms that may be as young as Middle Pennsylvanian and some contain forms that may be as old as Early Mississippian.'" (Copper Basin Group description)

Overlying the Copper Basin Group east of Mackay, Idaho, is a thick sequence called the White Knob limestone. Although the White Knob limestone (formerly the Brazer limestone) does not outcrop within the area covered by the present report, a description of this formation is included because it is a major unit just east of the report area (White Knob Mountains, Lemhi Range, etc.).

White Knob Limestone

Distribution. "Fairly early in the study of the Paleozoic rocks of south-central Idaho (Ross 1934b, p. 977-985) the name Brazer Limestone was applied to thick and widespread limestone beds in and south of southeastern Idaho. Diverse reasons have caused me to abandon this usage and, so far as south-central Idaho is concerned, to substitute the name White Knob Limestone (Ross, 1962) for beds there hitherto called Brazer Limestone and certain closely associated beds that are younger than the Brazer was supposed to be (Ross 1961a, p. 222-228). The name White Knob Limestone is derived from the White Knob Mountains in the Mackay quadrangle where the formation is well exposed.

The White Knob Limestone, of the revised nomenclature, is exposed in the eastern parts of T 9-11 N, R 20 E, and east of there (Ross 1937 (1938), p. 33-34). It is also exposed extensively at intervals throughout the Lost River Range and has been studied in northern parts of that range (Ross 1947, p. 1113-1117). In the Mackay quadrangle exposures are numerous but rather small. It is conspicuous in the White Knob Mountains but some of the most complete exposures in the quadrangle are in and just west of T 3 N, R 24 E. The outcrops in the central part of the quadrangle are the most westerly ones to which the name is applied. The formation is

widespread in the southeastern part of the Pioneer Mountains, southward to near the border of the Snake River Plain.

The limestone is widespread and thick in the Lemhi Range from the vicinity of Gilmore to almost the southeastern tip of the range. It is also present in hills between the Lemhi and Lost River ranges. It is probably the most abundant unit of Paleozoic age in the Beaverhead Mountains from T 17 N, R 25 E southeastward, although in Idaho little of it has been mapped. In parts of the Beaverhead Mountains the limestone rests directly on rocks of the Belt Series (Ross, Forrester 1947) but in places strata of early Paleozoic age intervene between the limestone and the Precambrian rocks (Scholten 1957, p. 153)."

Character. "In and near the Lost River Range the White Knob Limestone consists of a basal portion comprising several hundred feet of thin beds of limestone and chert, succeeded upwards by several thousand feet of thicker bedded limestone, which is everywhere the principal part of the formation. Both portions locally include sandy and argillaceous beds. The thin-bedded part of the formation is exposed in only a few places in the Mackay quadrangle. A similar basal component underlies the thick-bedded limestone in the Lemhi Range and in that the uppermost beds are younger than any known in the northwestern part of the Lost River Range and some or all may be absent there. These upper beds contain more varied rocks than the rest of the formation.

In the Borah Peak quadrangle (Ross 1947, p. 1111-1112) much of the thin-bedded basal component is calcareous but part is carbonaceous quartzite and some is chert. Sooty graphite is abundant in some beds.

The beds are commonly 2 to 6 inches thick and are crossed by numerous steep joints, commonly reddened by iron oxides. Consequently, the rock commonly forms nearly perpendicular cliffs which yield such abundant talus that they are locally almost buried. This feature renders the unit recognizable from a distance, and aids in distinguishing it from the softer, less calcareous Milligen Formation below. The basal unit in the Lost River Range is 300 to 500 feet thick.

In the Lemhi Range (Ross 1961a, p. 217-228) the corresponding rocks are similar to those in the Lost River Range but less conspicuous, in part because joint faces are not markedly iron stained. A few thick, fossiliferous, limestone beds are interspersed among the thin beds. Some of the impure limestone beds contain detrital quartz, plagioclase, and mica. Black chert laminae a fraction of an inch to a few inches thick are common

and introduce some difficulty in discriminating the lower part of the White Knob Limestone form the Milligen Formation which here also contains abundant chert. In most places in the southern part of the range, the thickness is over 1,000 feet but the unit thins and becomes less distinctive northward.

The thick-bedded main part of the White Knob Limestone has much the same characteristics wherever it occurs. The beds are one to several feet thick. The rock is bluish gray to almost black on fresh fracture and weathers light bluish gray. Chert nodules and laminae are common and in some beds are spaced only a few inches apart. Most of the rock is black on fresh fracture. Joints are less plentiful than in the thin-bedded basal unit. The limestone varies in detail but most of it is an aggregate of irregular calcite grains with subordinate grains of detrital quartz. In some beds the calcite grains are 0.005-0.01 mm in diameter and sizes range from this to over a millimeter. The quartz grains are all small, commonly 0.05-0.10 mm in diameter. Many have their original subangular outlines obscurred by added quartz. The matrix of most beds is fine-grained calcite but in a few the matrix is largely quartz. Most of the rock is pure calcium carbonate, apart from the chert nodules and laminae. Much of it is found on analysis to contain from 94 to over 98 percent calcium carbonate. Some contains small amounts of magnesium but beds that approximate dolomite in composition are rare. Sandstone, siltstone, and sandy limestone beds are present in several places but mostly in such small quantities that they are overlooked except where fairly detailed examinations are made. Exceptions to this are found in the southeastern part of the Bayhorse quadrangle and in places in the Mackay quadrangle in and near the eastern part of unsurveyed T 6 N, R 22 E; in these localities certain beds are sandy and even conglomeratic and their resistant outcrops are prominent features of the hillsides. Rocks of this character in and near unsurveyed sec. 14, T 6 N, R 22 E have been studied in detail by Betty Skipp (1961). The section she measured rests on argillaceous rocks of the Copper Basin Formation and has the upper part eroded away. The rocks of the White Knob Limestone (Brazer at the time the study was made) total 7,365 feet in thickness and have quartzose beds at intervals throughout but the conglomerate is in the upper 3,000 feet of the section within which they constitute 14 percent of the whole. Individual conglomerate beds could not be traced laterally for more than half a mile. The pebbles consist of chert, quartz, and quartzite and range up to 100 mm in diameter. Some of the conglomerate appears to have been deposited on beaches and some may result from turbidity currents.

While substantial amounts of the White Knob Limestone in the Borah Peak and Mackay quadrangles are included in

measured sections, the total thickness has nowhere been determined but may be roughly 10,000 feet for the thick-bedded part of the formation. This figure is based on consideration of the partial sections available. The total obviously exceeds 7,000 feet, as shown by the section reported by Betty Skipp and is thus materially greater than previous estimates (Ross 1934b, p. 978).

At the time that the southern part of the Lemhi Range was studied (Ross 1961a, p. 217-228) it became clear that the upper part of the sequence there included beds decidedly younger than the Late Mississippian age previously assigned to the bulk of the rocks then termed Brazer Limestone in south-central Idaho. In consequence the term 'Brazer limestone and related rocks' was used. All the rocks embraced in this provisional term are included in the White Knob Limestone as now defined. Future workers will split the White Knob Limestone into components wherever that proves to be possible.

The rocks in the upper part of the Upper Paleozoic sequence in the southern Lemhi Range have many resemblances to the thick-bedded unit in the Borah Peak and Mackay quadrangles but are more diversified than most of that unit is. They include quartzite, calcareous sandstone, sandy and argillaceous limestone, and oolite interspersed with more abundant limestone. Map measurements indicate a total thickness of 8,000 feet for the 'Brazer limestone and related beds.' Here, as in all other parts of south-central Idaho, no supercumbent rocks are known so the amount of limestone eroded off the top is impossible to determine. With due allowance for the large amount of limestone visible in parts of the range not yet mapped, a total thickness of 10,000 feet for the White Knob Limestone in the Lemhi Range is reasonable, although not positively established." (Ross description of White Knob Limestone.)

The Wood River Formation which overlies the Milligen Formation west of Copper Basin, has recently been subdivided by Wayne Hall (U.S.G.S., Menlo Park, California). Hall's stratigraphy is workable throughout much of the Bayhorse area and field mapping (in progress) has shown that the structure in the Paleozoic section throughout the region is far more complex than is shown on the Hailey and Challis AMS maps. Ross's 1962 description of the Wood River Formation is used in this report as Hall's stratigraphy is still undergoing refinement and is unpublished.

Wood River Formation

Distribution. "The Wood River Formation of present usage is essentially similar to the unit so named by Lindgren (1900, p. 89-90, 193-195) except that the lower part of the sequence he included in it is now termed the Milligen Formation. The Wood River Formation has been mapped in the Wood River (Umpleby, Westgate, Ross 1930, p. 29-34, pl. 1) and Bayhorse regions (Ross 1937 (1938), p. 36-39, pl. 1), and in the Borah Peak quadrangle (Ross 1947, p. 1117, pl. 1). Small exposures are present in the Mackay quadrangle. The small areas in the Borah Peak quadrangle originally assigned to the Wood River Formation are now regarded as belonging to the Copper Basin Formation and the same is true of beds in and near the southwestern part of the Bayhorse quadrangle.

By far the most extensive exposures are in and south of the Hailey quadrangle. Most of that quadrangle is underlain by the formation except that large areas are masked by the Challis Volcanics. There the Wood River Formation is widespread on both sides of the Big Wood River and the upper reaches of the Big Lost River. The exposures west of the Big Wood River extend into the Sawtooth quadrangle and the same or allied rocks reach to the border of the Snake River Plain south of both the Hailey and Sawtooth quadrangles.

The Wood River Formation is covered in the northwest part of the Hailey quadrangle by the Challis Volcanics but appears again on both sides of the southern end of Stanley Basin in the Sawtooth quadrangle. It extends in essential continuity from there northward through the Custer quadrangle into areas south and southeast of Bonanza. Beds which are in part surely, in part probably, of Pennsylvanian age are known at intervals from east of Darlington in the Lost River Range southward through the Arco Hills. Some of these resemble the Wood River Formation but, pending detailed study, it seems best to refer them provisionally to the Copper Basin Formation, a more inclusive unit."

Character. "Nearly everywhere in the Hailey quadrangle, where the unit was first recognized and named, a massive brownish quartizitic conglomerate is exposed at or near the base of the Wood River Formation. It averages 300-400 feet thick and is locally over 1,000 feet thick. In some places, such as the vicinity of Eagle Creek north of Ketchum, it is represented only by two or more lenticular conglomerate beds each rarely much more than 100 feet thick. At these localities conglomerate is not everywhere present near the base of the formation. The principal pebbles are black flint and gray quartzite, commonly 2-3 inches in diameter.

Above the conglomerate there are generally thick beds of dark-blue sandy fossiliferous limestone, which locally contains some conglomerate. It weathers blue-gray. The quartz sand grains in it average less than 0.3 mm in diameter and make up 20 percent of the rock. This limestone is so easily recognized that it is useful as a horizonmarker where the underlying conglomerate is absent.

Above the blue limestone is black impure limestone. Most of it is thin bedded and weathers to thin lavender-colored fragments and slabs. Some beds are comparatively thick. Northeast of Ketchum 1,200 feet of this limestone was measured.

The greater part of the Wood River Formation is calcareous and quartzitic, thick-bedded, crossbedded sandstone with shaly layers. This unit ranges up to at least 2,800 feet in thickness. The more siliceous varieties are light gray and rather fine. Many of the original detrital grains have been secondarily enlarged; some mica and other impurities are commonly present. The rock containing calcareous material is dark and generally weathers rusty. Such rock consists essentially of angular to subangular quartz grains with minor amounts of feldspar, muscovite, magnetite, carbonaceous material, and zircon in a fine-grained calcareous cement. A sample of calcerous sandstone containing Fusulina was analyzed by W. C. Wheeler, of the U.S. Geological Survey, with the following result:

Analysis of blue calcareous sandstone from the Wood River Formation

Insoluble	74.32 percent
MgO	1.85 percent
CaO	11.33 percent
Loss on ignition	11.68 percent
	99.18

Interbedded with these calcareous and siliceous sandstones are darker, medium-bedded rocks of distinctly dolomitic character.

A facies of the Wood River Formation, slightly different from the others, is found near the southwest corner of the Hailey quadrangle. The rock is a blue-black, fine-grained siliceous and carbonaceous limestone, weathering dark gray. It is commonly laminated, alternating beds varying in the proportion of siliceous and carbonaceous matter. Quartz grains scattered through the limestone range up to 0.1 mm in diameter. In some

places the bands of more siliceous material are irregular and appear to be flintlike aggregates in the rock.

In the upper part of the formation the typical rock is a dark, fine-grained calcareous sandstone, which bears on the fresh surface a very characteristic bluish and brownish-black mottling. It is even bedded; the beds in places are as much as 5 feet thick, commonly separated by thin bands of shaly rock which bring out the bedding in a striking manner. The rock weathers generally to a rusty brown, and in some places to a brownish gray or gray.

A partial analysis of this rock by W. C. Wheeler gave the following result:

Analysis of calcareous sandstone from the Wood River Formation

Insoluble	69.85 percent
MgO	5.66 percent
CaO	9.24 percent
Loss on ignition	14.78 percent
	99.53

Interbedded in these calcareous sandstones are beds of black chert, which occur in massive layers as much as 10 feet thick, breaking into bands 1-2 inches and in some places as much as 6 inches thick. Blue-black limestone, weathering blue gray and containing bands and lenses of black chert, is interbedded at a few horizons. Beds of quartzitic sandstone are intercalated in places. The maximum observed thickness of this unit exceeds 2,500 feet.

No single continuous and complete section of the Wood River Formation has been found. In the Hailey quadrangle, separate sections, which differ from one another in lithologic character and so are believed not to overlap in any considerable degree, give a total of 7,700 feet. In all probability these measured sections do not represent the whole of the formation; 8,000 feet is taken as a minimum thickness for the Wood River Formation in this part of the region.

In the northern part of the Sawtooth quadrangle and the southern part of the Custer quadrangle, the rocks assigned to the Wood River Formation are essentially identical in general lithologic character with those in the Hailey quadrangle. Close to some of the igneous contacts here and to an increasing extent northward the strata have, however, undergone intense igneous metamorphism. This has so changed their appearance that it is only by tracing the rocks through the progressive changes that the more thoroughly altered strata can be recognized as belonging

to the formation. The less metamorphosed rocks are mainly quartzite. They vary in color from light gray to nearly black, with increase in impurities, especially carbonaceous matter. Many are somewhat calcareous and the more limy beds tend to weather rusty.

Limestone beds are distinctly subordinate in amount in the southern Custer and northern Sawtooth quadrangles. It appears, however, that the more conspicuous metamorphism farther north results to some degree from an increase in easily replaceable limestone beds. Much of the quartzite is somewhat argillaceous, but shaly material is nowhere abundant and is commonly confined to thin partings spaced several feet apart between the quartzitic beds. Nearly everywhere outcrops of this formation have a massive appearance, with readily visible partings 2-5 feet apart. Finer banding and crossbedding are, however, almost invariably perceptible on close inspection.

Conglomerate occurs in several localities at or near the base of the formation here but is not as conspicuous or constant a feature as in the Hailey quadrangle. The more abundant pebbles are quartz and quartzite, but the matrix is somewhat more impure and the rock consequently less resistant to weathering than in that region.

The outstanding feature of the metamorphism is the conversion of normally somber beds into dazzling white rocks. This results in part from elimination of organic matter, but mainly from the abundant development of white wollastonite and diopside. In those limestone beds not thoroughly replaced by silicates, the recrystallization of the calcite has lightened the color of the rock.

In and near the eastern part of T 8 N, R 20 E, in the Bayhorse quadrangle, a small area has been mapped as underlain by rocks belonging to the Wood River Formation because they are coextensive with strata so assigned in the adjoining part of the Hailey quadrangle. These beds comprise an alternating sequence of conglomerate, grit, and carbonaceous argillite. There is no essential lithologic difference between certain of the rocks that were mapped as components of the Wood River Formation and some nearby that have been mapped as Milligen because they lie beneath the Brazer (now White Knob) Limestone and above the Grand View Dolomite. The relations between the supposed Wood River beds and those of supposed Milligen age are masked by a cover of Tertiary conglomerate.

The mapping of the Wood River Formation in the Wood River region extended south of the Hailey quadrangle as far as the vicinity of Bellevue. Later a stratigraphic study, unsupported by mapping, was made near Bellevue (Bostwick

1955). In this study some 12,000 feet of beds were examined. There may be some duplication by faulting. Bostwick evidently thought his section began at the base of the Wood River Formation, and that the lower third is of Pennsylvanian age and the rest of Permian age. The implication is that the Pennsylvanian beds total about 8,000 feet in the Hailey quadrangle and about half that thickness near Bellevue. This seems unlikely but the apparent inconsistency cannot be cleared up with the information at hand.

Bostwick's (1955, p. 949) descriptions of his collections indicate that he found a basal conglomerate with thin-bedded limestone 20 feet above it and fossiliferous limestone and sandy limestone at frequent intervals throughout the rest of his section. A few of his collections were from calcareous sandstone. He describes the basal conglomerate as made up of angular to well-rounded chert and quartzitic pebbles averaging an inch in diameter, with abundant interstitial sand. Small lenses of sandstone and sandy limestone are common in the conglomerate. Sandy shale is interbedded with limestone above the conglomerate. Sandstone, abundant higher in the sequence, consists mainly of quartz with interstitial calcite; some contains a little feldspar.

Hewett (in Umpleby, Westgate, Ross 1930, p. 210-211) gives a section on slopes southeast of Bellevue that differs in some details from Bostwick's descriptions. This section is reproduced below.

Sandstone, calcareous, bluish gray	600+ feet
Limestone, dark gray, no fossils; contains many white calcite veinlets, persistent	80 feet
Sandstone, calcareous, gray	200 feet
Sandstone, buff, not calcareous	150 feet
Sandstone, coarse, calcareous, gray	240 feet
Shale, buff	50 feet
Sandstone, calcareous, bluish gray, with beds of buff sandy shale	300 feet
Sandstone, calcareous, bluish gray; veins of black calcite	50 feet
Shale, sandy, buff	20 feet
Limestone, siliceous, bluish gray, fossiliferous	35 feet
Limestone, white	12 feet
Conglomerate, containing many lenses in which the pebbles range from 1-4 inches in diameter; fossils locally abundant in sandy layers	300 feet
	2,037

(Ross's description of Wood River Formation.)"

A BRIEF GEOLOGIC HISTORY OF PALEOZOIC SEDIMENTATION IN SECTION II

The interpretation of the Paleozoic history of central Idaho appears to be directly related to two major orogenic events, the Antler orogeny (Late Devonian to Early Mississippian) and the Laramide orogeny (Late Cretaeceous to Early Tertiary). Roberts et. al. (1965) have given a composite picture of the Paleozoic history of Utah, Nevada, and Central Idaho.

Figure 14 is a panel diagram showing the relative thicknesses of Middle Paleozoic sediments in Idaho and Southwestern Montana (Churkin, 1962). Relatively thin carbonates (miogeosynclinal sediments) are located in the east while a thick clastic pile (eugeosynclinal sediments) lies to the west. Churkin interprets this sequence as having an eastern (cratonic) source area. The miogeosyncline and eugeosyncline were part of the Cordilleron geosyncline which was affected by the Antler orogeny (Figure 15 shows the extent of the Antler orogenic belt). This orogeny thrust the eugeosynclinal clastic rocks onto the miogeosynclinal carbonate sequence as is shown in Figure 14.

The major effect of the Antler orogeny was twofold; eugeosynclinal rocks were thrust on miogeosynclinal rocks and regional uplift of the entire Antler belt occurred. This new "high" provided a source area for clastic sediments which were shed eastwards (Copper Basin Sequence) and westwards (Milligen and Wood River Formation). This was the first time a western source area was providing sediment (coarse clastics, shales and volcanic material) to the geosyncline.

Between the close of the Antler orogenic episode (Late Paleozoic) and the beginning of the Laramide orogeny (Late Mesozoic) sedimentation continued from the western source area providing material east and west. Clastic calcareous sediments were deposited to the west (Wood River Formation) while limestones formed to the east (White Knob Limestones). During the Laramide orogeny the western sediments were thrust eastwards. Emplacement of the Idaho batholith at this time caused more structural complexities and probably more thrusting. This is a broad interpretation of what the geology of Central Idaho reveals. It will undoubtedly be changed as better mapping becomes available.

Tertiary Volcanics and Related Sediments

Ross first described the Challis volcanics in 1927 (updated 1934c). His description of the volcanics (as with the Paleozoics) is sufficient for the map scale used in this report. Some confusion was brought on by Anderson (1961) who renamed the basic Challis units. Ross's three major units of the Challis volcanics are a basal latite-andesite member, (which in places is underlain by a conglomerate which Anderson called the Beaverhead conglomerate), the middle Germer tuffaceous member and upper Yankee Fork rhyolite. The rhyolite member is covered in the northeastern part of Section II by a sequence of sediments, in part volcanic, which are most likely lake deposits. The volcanic units and their map symbols are shown on the legend in the map folio. Much of the Challis volcanics have been

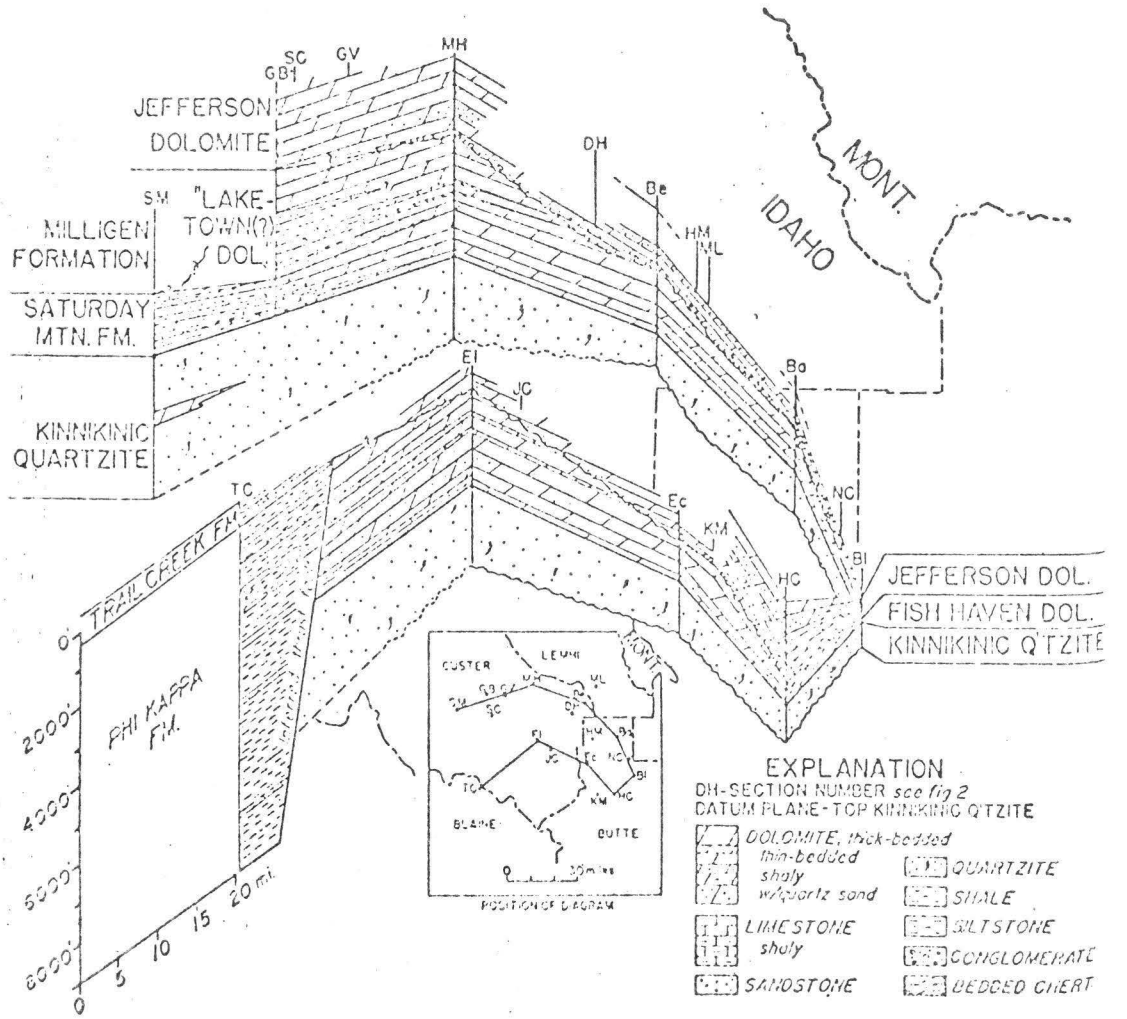


Figure 14. Panel diagram of Middle Paleozoic strata (Churkin, 1962)

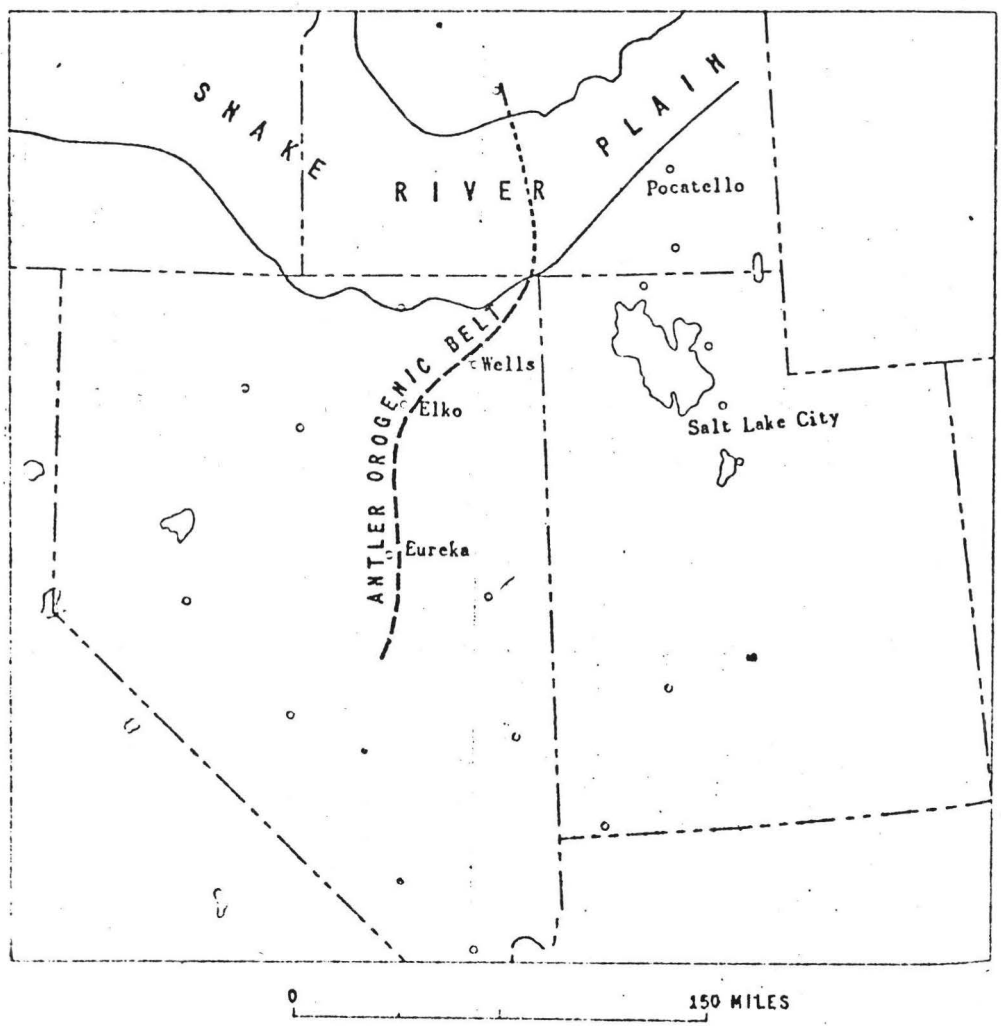


Figure 15. Map showing eastern limit of Antler orogenic belt in Late Devonian to Early Mississippian time (Roberts, et. al., 1965)

mapped as undifferentiated Challis volcanics.

Challis Volcanics

Definition and Subdivision

"The most conspicuous and widespread assemblage of dominantly volcanic rocks in the eastern part of south-central Idaho has been known for a long time as the Challis Volcanics. The original definition (Ross 1927, p. 79; 1934a, p. 46) was necessarily broad because so little detailed information was available. The term Challis Volcanics is now restricted (Ross 1961b) to dominantly volcanic strata of early Tertiary age within the part of central Idaho north of the Snake River Plain and south of the westward-flowing segment of the Salmon River (near latitude 45° 30'). Thus, the Challis Volcanics represent the lower part of the Tertiary deposits of south-central Idaho. As restricted in the paper just cited they may range in age from Eocene into, at the most, early Miocene. Probably most of the formation is Oligocene.

In most areas in which the Challis Volcanics have been studied in any detail, local subdivisions are recognizable. Many of these have so many resemblances to each other that one is tempted to make correlations. When study is farther advanced than it now is, some such correlations are likely to prove to be justified. At present any that are made must be regarded as tentative.

The part of the Challis Volcanics that first attracted attention is that composed mainly of flow rocks of intermediate composition. Apart from the fillings of certain valleys and basins, it is the dominant component of the formation. It is referred to here as the latite-andesite member. This member varies in details of composition and other features from place to place. In the Lemhi quadrangle, Anderson (1961, p. 65-67) called the member the Cheney Volcanics. In that area the rocks are lighter colored and in part more silicic than is common in the latite-andesite member. The extent to which these features may correspond to stratigraphic differences remains unknown.

In numerous localities conglomerate is present in mappable amount at, or close to, the base of the Challis Volcanics. Even though conglomerate may not have been laid down at the same time everywhere, it is a stratigraphic feature of much interest. In the North Fork quadrangle Anderson (1959, p. 21-22) calls it the Kriley Formation and in the Lemhi quadrangle (Anderson 1961, p. 30-31) he refers to it as the Beaverhead conglomerate.

Thick assemblages of dominantly tuffaceous rocks are present here and there, mostly as fillings of depressions. Commonly flows are intercalated in the tuffaceous beds. One of the first of the tuffaceous assemblages to be mapped is called the Germer Tuffaceous Member (Ross 1937 (1938), p. 53-58). Similar assemblages are present near Salmon and in other localities. In the Lemhi quadrangle Anderson (1961, p. 48-52) calls one such assemblage the Yearian Volcanics. He notes its resemblance to the Germer Tuffaceous Member. In his work near Salmon (Anderson 1956, p. 22-24) he spoke of a tuffaceous unit. His descriptions suggest that the tuffaceous unit near Salmon and the Yearian Volcanics of the Lemhi quadrangle have much in common.

Presumably most of the thick tuffaceous assemblages in south-central Idaho were never interconnected and therefore their mutual relations cannot be established by direct tracing of one into another. Perhaps the tuffaceous beds scattered through the latite-andesite member will furnish clues after refined mapping is accomplished. Discovery of fossils that could be closely dated or of some other means of fixing the age of the beds would be of great assistance but, at the moment, data of this kind are inconclusive. It is a natural assumption that the large eruptions of ash that produced thick tuffaceous assemblages may have taken place at about the same stage in Challis volcanism but the extent to which this assumption may be correct is for future work to determine.

Flows of basalt and calcic andesite have been mapped in various places in the region and apparently through a wide range in stratigraphic position within the Challis Volcanics. Some of the most conspicuous of these are in the Germer Tuffaceous Member; Anderson (1956, p. 26-27; 1957, p. 32-33; 1961, p. 48-53) reports similar rocks near Salmon and in the valley of the Lemhi River above that town. He singles out one assemblage composed mainly of calcic lava as the Kadletz Volcanics, overlying his tuffaceous Yearian Volcanics. His Kadletz Volcanics may differ little from other flows intercalated in the tuffaceous assemblages, even though it appears to rest on an erosion surface. On the basis of present information the fact that a particular flow or aggregate of flows is of calcic composition cannot be taken as of stratigraphic significance, except that most basalt that has been mapped is associated with tuff.

Recent work by Anderson (1956, p. 28-31; 1957, p. 15-19; 1959, p. 21-28; 1961, p. 31-35) has resulted in emphasis on the presence in areas near Salmon of assemblages in which sedimentary beds predominate. He regards these as distinct from the Challis Volcanics but they have so many features in common with beds in that formation that provisionally they are regarded as members of the

Challis Volcanics. Fossils that in the past have been used as aids in dating the volcanics came from one of the sedimentary units. The scanty evidence at hand indicates that all of the units named by Anderson may fall within the age range of the Challis Volcanics as now restricted. If any are demonstrated by future work to be separated from the Challis by an unconformity of regional significance or to be materially younger than that formation, they should then be excluded from it.

Table 1, based on interpretation of Anderson's reports, shows the interrelations of units he discusses and names. As he does not everywhere map some components separately, equivalences can be shown only approximately.

One other recognized component of the Challis Volcanics requires mention. This is the Yankee Fork Rhyolite, distinguished so far only in and close to the Casto and Bayhorse quadrangles (Ross 1934a, p. 52; 1937 (1938), p. 59-62). This is a distinctive rock, found mostly in association with the Germer Tuffaceous Member. Perhaps it is a product of special conditions confined essentially to areas where it is already known, but the Challis Volcanics contain numerous silicic rocks and some of these may prove, on further study, to be more closely akin to the Yankee Fork Rhyolite than is now realized.

The Challis Volcanics cover a larger aggregate area than any other single stratified formation within south-central Idaho. Erosion and deformation have separated the exposures of the formation into areas whose combined extent is nearly 3,000 square miles, whereas the present exposures of the Belt Series, the next most extensive unit, aggregate less than 2,000 square miles. The distribution of the erosion remnants of the Challis Volcanics shows that, with the possible exception of some of the highest peaks, the formation originally covered almost all of Lemhi and Custer counties, eastern Valley County, all of Blaine County, with the possible exception of its southern border, and parts of Camas and Butte counties. The principal existing exposures extend, with interruptions, from the drainage basin of the Yankee Fork of the Salmon River northeastward to the northern part of the Lemhi Range. Another large mass reaches from near the Middle Fork of the Salmon to the vicinity of Big Creek (Edwardsburg on some maps). Ramifying exposures cover much of the country from the Salmon River near Clayton south and southeast to the border of the Snake River Plain, including areas in the Pioneer, White Knob, and Sawtooth Mountains. Scattered areas of volcanic rocks in the Lost River Mountains and the Lemhi Range are also known."

Table 1. Tentative correlation by C. P. Ross of components of the Challis Volcanics in the region near Salmon
Based on reports by A. L. Anderson (1956, 1957, 1959, 1961)

North Fork quadrangle	Salmon quadrangle	Baker quadrangle	Lemhi quadrangle
Kirtley Formation		Carmen Formation	Kirtley Formation
Geertson Formation		Geertson Formation	Geertson Formation
Kenney Formation	Carmen Formation	Kenney Formation	Kenney Formation
Flows and pyroclastic rocks. The latter are abundant especially in the upper part of the unit.	Tuffaceous beds with some flows	Largely tuffaceous beds	Kadletz Volcanics
	Basalt calcic andesite member	Intercalated basalt - calcic andesite member mapped separately because of interbedding	Yearian Volcanics
	Latite-andesite member	Latite-andesite member	Cheney Volcanics
Kriley Formation			Beaverhead Conglomerate

Summary Statement

"The character and succession of flows and associated strata vary from place to place but certain components are sufficiently uniform throughout the region to aid in description. These are, approximately in ascending stratigraphic order, the latite-andesite member, basalt and related flows, the Germer Tuffaceous Member and its approximate equivalents and the Yankee Fork Rhyolite. The calcic rocks have a wide range in stratigraphic position. At present, the formal name Germer is applied only in the Bayhorse region, but in numerous places tuffaceous and sedimentary rocks somewhat similar in character and stratigraphic relations to the Germer are known. The Yankee Fork Rhyolite is not strictly uniform either in character or in stratigraphic position. It has been mapped in the Casto quadrangle and the Bayhorse region and undoubtedly extends beyound these limits. Rhyolitic rocks are known in many places in south-central Idaho and neighboring regions. Some of these may differ only in details from the Yankee Fork Rhyolite. The relations of others are not yet determined. Wherever adequately displayed, the Challis Volcanics are over 2,500 feet thick; locally the maximum thickness appears to exceed 5,000 feet."

Latite-Andesite Member

"The latite-andesite member of the Challis Volcanics has been studied by me more thoroughly in the Casto, Bayhorse, Hailey, and Mackay quadrangles than elsewhere. The entire existing part of the formation in the Mackay quadrangle belongs to this member and so does most of the formation along the lower reaches of the Yankee Fork of the Salmon River (Anderson 1949, p. 9-10). The flows in the northern part of the Lemhi Range and in and near the eastern Salmon River Mountains close to Salmon are similar (Umpleby 1913a, p. 47-48; Anderson 1956, p. 18-27; 1957, p. 30-33; 1959, p. 22-25; 1961, p. 43-53). The unit consists mainly of flows and flow-breccias with the composition of quartz latite, latite, dacite, and andesite, with subordinate amounts of rhyolitic tuff and augite andesite and basalt locally intercalated in them. Much of the rock in the Mackay and Hailey quadrangles is more calcic than the corresponding part of the formation farther north, and the corresponding rocks along the Lemhi River may be relatively silicic on the average.

The name latite-andesite member is elliptic, mainly in the omission of reference to the quartz that is present in many of its components. It was first used in the report on the Bayhorse region (Ross 1937 (1938), p. 50-53) but corresponds approximately to the term "latite and hornblende andesite" that was applied in the work

in the Wood River region (Umpleby, Westgate, Ross 1930, p. 55-56). Rocks that correspond fairly closely to those of the latite- andesite member are present in nearly all areas in which the Challis Volcanics are exposed and in many places constitute much or all of the formation.

The latite and quartz latite flows are commonly some shade of purple or lavender. Many are gray and some are brownish and reddish. The more silicic ones are light pink and lavender. The rocks commonly have feldspar phenocrysts and contain conspicuous biotite and hornblende. They consist essentially of plagioclase (commonly oligoclase), potash feldspar, quartz, biotite, and hornblende. The accessory minerals are magnetite, apatite, and zircon. In most places the rocks are fairly fresh but many of them contain sericite, chlorite, calcite, and iron oxides as alteration products. Some of the more silicic flows are microspherulitic. Welded tuff may be present locally and tuff and coarser pyroclastic rocks are included in the unit in many places. Probably the majority of the flows in the latite-andesite member contain a sufficient proportion of a potash feldspar to be regarded as latitic, but in few of these is the potash feldspar readily recognizable, even under the microscope. The three sets of analyses tabulated below all record potassium, presumably in feldspar. Free quartz is present in many but is rarely conspicuous.

The latite-andesite member also contains numerous flows of andesitic composition. Lava of this sort is somewhat more abundant in parts of the Hailey quadrangle and in most of the Bayhorse quadrangle than it is in the Custer and Casto quadrangles. As a consequence, the member in the Hailey and Bayhorse quadrangles tends to be relatively somber and dark. Feldspar phenocrysts are less prominent in the more andesitic flows than in the latitic flows. Some of the andesitic rocks are so fine grained that the components cannot be determined even under the microscope. In the coarser andesitic varieties the phenocrysts are andesine or labradorite, hornblende and rarely biotite, and altered augite, or locally, hypersthene. The groundmass is largely fine grained feldspar, with some quartz, which may be of secondary origin. The secondary minerals--chlorite, sericite, calcite, zircon, and iron oxides--are abundant.

Throughout the region underlain by the latite-andesite member, but especially close to its top in parts of the Custer and Casto quadrangles and in the Thunder Mountain district, there are devitrified, non-porphyritic, and generally spherulitic rhyolite flows with greenish, reddish, purplish and other colors. A small amount of obsidian is associated with the sperulitic rocks, and

in related flows. Some of the rhyolitic and silicic latitic rocks may be welded tuffs.

The rocks of the latite-andesite member in the general vicinity of Salmon (Anderson 1956, p. 18-27; 1957, p. 30-33; 1959, p. 22-25; 1961, p. 43-53) are so similar to those described above as to lead Anderson to apply the same member designation to those around Salmon. The strata range in color from black to pink, lavender and nearly white. Most are porphyritic. They include andesite, latite, quartz, latite, dacite, and tuff. The breccias that are conspicuous in some of the area above described are not reported by Anderson but he notes that in the areas he studied some of the flow-like rocks are actually welded tuff. Flows of quartz latite are topped by black obsidian, which is absent from the welded tuff. Obsidian is present in most other areas in south-central Idaho where the Challis Volcanics remain but is rarely abundant or conspicuous. The latite-andesite member near Salmon, as elsewhere, includes some silicic tuff.

Anderson's descriptions indicate that within most parts of the areas he mapped, flows and welded-tuff units that resemble flows are subordinate to tuff and sedimentary rocks. For this reason he does not speak of the latite-andesite member either near Baker or along the Salmon River from Carmen to North Fork. However, in the mountains near Baker and farther up the valley of the Lemhi River there are thick masses of flows and flow-like rocks that are at least closely akin to the latite-andesite member of other areas. A small mass of similar rock is exposed near the North Fork of the Salmon River, mostly north of the area Anderson mapped.

In the Lemhi quadrangle, Anderson (1961, p. 43-53) divided the part of the Challis Volcanics in which flows predominate into three units. In ascending order, he named these the Cheney Volcanics, Yearian Volcanics, and Kadletz Volcanics. In consequence of this subdivision, he proposes to raise the Challis Volcanics to group rank. Since that unit was first described it has been appreciated that it contained substantial subdivisions susceptible to being mapped. On this ground the unit might from the first have been regarded as a group. The naming and mapping of three subdivisions in part of the unit in one locality does not change the status of the assemblage in this respect. Whether it is labelled a group or a formation, the Challis Volcanics is a diversified assemblage whose components vary from place to place. Anderson's work in the valleys of the Salmon and Lemhi rivers has emphasized the diversified character of the assemblage.

Anderson notes that his Cheney Volcanics are widely distributed in the south-western part of the Lemhi quadrangle and are less abundant in the northeastern part. He remarks that they are plentiful west and northwest of the Lemhi quadrangle and are the dominant volcanic unit in the Salmon area. Thus, they constitute a large element, if not all, of the latite-andesite member of the Challis Volcanics, an informal subdivision of long standing.

The Cheney Volcanics are light-colored, bright-appearing, gray, pink and lavender (in part in deep shades) latitic and andesitic flows and flow breccias, with subordinate quantities of latitic and rhyolitic welded tuff. The flow rocks are fine-grained, in part glassy, and have small phenocrysts. The rocks include hornblende andesite, hornblende- biotite andesite, hypersthene-biotite andesite, biotite andesite, hornblendebiotite latite, quartz-latite and rhyolite. The proportions of these different rocks vary from place to place. The thickness is over 2,000 feet.

The Yearian Volcanics are the most widely exposed unit in the Lemhi quadrangle and are especially abundant in the Lemhi Range. They consist mainly of light-colored, silicic tuff and related pyroclastic rocks, with only minor intercalations of andesitic and basaltic flows. Most of the tuff is postulated by Anderson to have accumulated by ash showers from many local vents. Some tuff is water-sorted and a little tuffaceous sandstone and conglomerate is included. In some places flows constitute much of the formation. They are dark-gray to black rock commonly with few phenocrysts. They range from calcic andesite to basalt, depending on the feldspar contained in them. The original thickness of the unit was as much as 3,000 feet, at least locally. Anderson remarks that his Yearian Volcanics have much in common with the Germer Tuffaceous Member of the Bayhorse quadrangle."

Basalt and Related Flows

"Flows originally described under the heading "Augite, andesite and basalt" (Umpleby, Westgate, Ross 1930, p. 55) apparently occur at several horizons in the Wood River region. Westgate concluded that the majority lay at the base of the Tertiary sequence. However, the two largest individual masses, which lie along the northern border of the Hailey quadrangle, are above the andesitic and latitic flows of the locality as is shown both by Westgate's map and by studies in adjoining parts of the Bayhorse quadrangle. Flows of calcic composition are probably present in parts of the areas cited above as underlain mainly by the latite-andesite member.

In the Hailey quadrangle the rocks of the sort here described consist mainly of augite andesite, basalt being rare. Most of them are grayish or brownish black, augite andesite and weather to a deep-red soil. The phenocrysts are labradorite and colorless augite. The groundmass is very fine grained and consists of andesine, augite, and magnetite. Typical olivine-basalt without feldspar phenocrysts is present near the Red Elephant mine, in the southwestern part of the region and in a few other places. The basalt and andesite are on the whole remarkably fresh.

In the Bayhorse quadrangle, in addition to large and thick masses of augite andesite and basalt like those in the Hailey quadrangle, flows of basalt are intercalated in the tuffaceous beds of the Germer member in many places. In most such flows phenocrysts make up an average of about 20 percent of the whole. They consist of augite and labradorite in varying proportions. Biotite is occasionally present in small amount. Olivine is absent or inconspicuous.

North and northwest of the Bayhorse quadrangle calcic flows are much more rare and in most localities are absent. In the Thunder Mountain district a number of basalt flows of restricted lateral extent are associated with the rhyolite and tuff, especially just above the Dewey mine. Some of these rest on the clastic rocks and merge with dikes which cut them (Shenon, Ross 1936, p. 11).

Basaltic flows (Umpleby 1913a, p. 48) are prominent in the hills just below the confluence of the Pahsimeroi and Salmon rivers and also in areas along the Lemhi River. They are fine-grained, bluish-black rocks of normal basaltic composition, except that the proportion of olivine is markedly variable.

Anderson's Kadletz Volcanics (Anderson 1961, p. 52-53) evidently belong to the basaltic component of the Challis Volcanics whose stratigraphic relations vary widely in different parts of Idaho. He notes that the Kadletz Volcanics are the least widespread of the three volcanic units he distinguished in the Lemhi quadrangle. He says calcic flows extend into the areas around Baker and Salmon where he termed them the basalt-calcic andesite member (Anderson 1957, p. 32-33; 1956, p. 26). The Kadletz Volcanics consist mainly of basalt but, north of the Lemhi quadrangle, include calcic andesite and, at the top, some latite."

Germer Tuffaceous Member

"The Challis Volcanics in many places include beds of tuff and of sedimentary rocks that in varied degree contain tuffaceous material. Of these, the Germer Tuffaceous Member was the first to be formally named (Ross

1937 (1938), p. 49, 53-58). For convenience, in the absence of other names and because of uncertainties as to exact stratigraphic relations, the term "Germer tuffaceous member and related strata" was employed on the geologic map of Idaho (Ross, Forrester 1947) and is retained here. This term refers to assemblages of dominantly clastic rocks in the formation, by no means necessarily stratigraphically equivalent throughout.

The Germer Tuffaceous Member takes its name from Germer Basin south of the Salmon River and about 13 miles southwest of Challis. It has been mapped in this and neighboring parts of the Bayhorse quadrangle and in Round Valley near Challis. Application of the formal name in other localities is of doubtful validity at present. The name has been used (Anderson 1949, p. 10) for rocks in the northern part of the Custer quadrangle but these include much non-tuffaceous material. The tuffaceous beds in the upper part of the Challis Volcanics in the Hailey quadrangle, the northern part of the Custer quadrangle, the Casto quadrangle, and northwestward across Monumental Creek almost to Big Creek have many similarities. Some of those near Salmon and along the Lemhi River are also similar but here some contain large proportions of sedimentary material. Dominantly sedimentary deposits are also known in the Mackay quadrangle and east of Stanley.

The Challis Volcanics contain clastic beds in many places in quantities too small to show on existing maps. These probably are present in one locality or another, throughout the formation. Most of them are lithologically akin to the main parts of the Germer Tuffaceous Member, described below. South-central Idaho also contains clastic rocks younger than the Challis Volcanics. Few of these have yet yielded diagnostic fossils, and the distinction between them and clastic components of the Challis Volcanics cannot everywhere be made positively at the present time.

By far the greater part of the Germer Tuffaceous Member in and near its type locality consists of medium to fine-grained, pale-orange or buff and light-brown rocks composed largely of the products of explosive volcanism. Much of it shows bedding and water-sorting to a greater or less extent; much of this clearly water-lain rock is in beds a few inches to a few feet thick, but in many bluffs tuff masses a foot to several score feet thick exhibit little or no bedding.

The most abundant variety of the tuff is composed mainly of crystals of igneous minerals, commonly about a millimeter in maximum diameter, embedded in a matrix of similar character but with many of the grains only about 0.02 mm in diameter. The matrix is partially silicified.

Most fragments in it are sharply angular. Among the larger fragments oligoclase crystals are abundant. Some of these are almost perfect in form but many are broken. Quartz, altered feldspar, biotite, hornblende, and bits of lava are also discernible. Some of the altered feldspar is probably orthoclase but, as in much of the latite-andesite member, few grains of potash feldspar have been positively identified. In some beds, especially those close to lava flows, fragments of lava of several kinds predominate and fragments of crystals are subordinate. These beds are silicified, in part so much so as to be flint-like in appearance.

A few of the tuff beds are distinctly harder and denser than the average, although not perceptibly silicified. These contain fragments of minerals but consist mainly of delicate glass shards with an average length of about 0.1 mm. These shards must be products of ash fall, with essentially no later transportation in water. Some beds of similar appearance to these are welded tuff, composed of pumice in which the glassy fragments erupted were so hot when they fell on the ground that they flowed and welded together into a mass of glassy material with exceedingly irregular and discontinuous lamination.

Tuff has been quarried for building stone at Challis and structures made of it have proved to be resistant to weathering if adequately protected from frost action at ground level. Some of these buildings have survived fires that did wide damage in the town. Behre (1930) gives a detailed section of 330 feet of the tuff in the quarry at Challis. His petrographic description is essentially like the more generalized one given above. Behre records results of tests of compressive strength, absorption, and weathering made at the U.S. Bureau of Standards. The welded tuff contains a few microliths of feldspar and biotite, and sparse crystals of sodic plagioclase and quartz.

The tuff that exhibits good bedding contains some water-worn quartz fragments. As the proportion of such material increases the tuff grades into tuffaceous sandstone but sandstone appears to be distinctly subordinate throughout the type locality of the Germer Member. Thin layers of silt and clay occur locally but are rarely either thick or abundant. The better preserved plant remains are generally found in such rock. In a few places thin seams of impure lignite are present but none of these are of economic value as fuel. Wood fragments, logs, stumps, and tree leaves are widely distributed in the Germer Tuffaceous Member. Much of the wood was worn by water before it was buried, but the stumps of a few trees, some of them large, were still in the process of growth. All of the wood has been extensively altered, mainly by silicification.

Conglomerate is conspicuous in several localities, especially near the base of the member. Most of it has a matrix of tuffaceous sandstone, but some of it has a matrix composed mainly of grains of quartz and quartzite in a siliceous cement. The pebbles and boulders in the conglomerate range in diameter from a quarter of an inch to several feet. Conglomerate with a tuffaceous matrix contains pebbles of volcanic rocks mingled with many from the Paleozoic rocks. In the conglomerate with siliceous cement volcanic rocks are scarce and the pebbles are more rounded than in those with tuffaceous matrix. A few beds contain pebbles of granitic rocks that must have been transported long distances from their probable sources.

The Germer Tuffaceous Member rests on an irregular surface carved on pre-Tertiary rocks and on its margins interfingers with flows belonging to the latite-andesite member. Flows of rhyolite and basalt are present in the tuffaceous beds. The thickness of the tuffaceous member is variable but probably nowhere much exceeds 1,000 feet.

Beds that are so similar to the Germer Tuffaceous Member that Anderson (1949, p. 10) used that name for them, are plentiful in the northern part of the Custer quadrangle. They are not coextensive with the Germer beds of the type locality and so much of the work so far done in the Custer quadrangle is of reconnaissance character that exact stratigraphic relations are in doubt.

In the area near Bonanza, Anderson says the tuffaceous beds and the flows interleaved with them are over 3,000 feet thick. He says much of the tuff is fine to medium grained, pale buff to light brown rock, the product of ash showers. Much is bedded and water-sorted. Silty and shaly beds are present locally. Much of the tuff whether sorted or unsorted is composed of fragments of feldspar and other minerals, but some of it contains fragments of glass; other beds are composed largely of welded pumice. The two analyses below will give an idea of the chemical composition of the tuff in and close to the Custer quadrangle. One of the principal differences between the Germer Tuffaceous Member and the otherwise similar beds in the northern part of the Custer quadrangle is in the large proportion in the latter area of lava flows similar to those in the latite-andesite member."

Beds other than the Germer Tuffaceous Member

"Tuffaceous and sedimentary rocks that to a greater or lesser extent are lithologically like the components of the Germer Tuffaceous Member are intercalated in nearly

all of the Challis Volcanics. The characteristics of the principal assemblages of this sort are outlined below.

Prospecting for uranium in and near T 11 N, R 14 E, east of Stanley (Kern 1959, p. 9-10), has uncovered several areas of sedimentary rocks capped by flows and tuffs that are thought to belong to the latite-andesite member. The sedimentary rocks constitute the base of the Challis Volcanics locally. They consist largely of tuffaceous sandstone and siltstone but include lenses of impure lignite and of fluviatile, arkosic conglomerate. The sedimentary beds reach thicknesses of over 100 feet but the total is nowhere exposed. The relations between these isolated exposures and the tuffaceous beds northeast of them are not known.

Tuffaceous beds are plentiful north of the Custer quadrangle but are so interspersed with lava that they have not been mapped separately (Ross 1934a, p. 45-52, pl. 1). In composition they closely resemble those in the Custer quadrangle. In and east of the Casto quadrangle the thickness is 1,500-2,000 feet where most abundant but farther west in the Casto quadrangle thicknesses range from zero to a few hundred feet.

West of the northern part of the Casto quadrangle the Challis Volcanics contain large amounts of tuffaceous beds. Near Monumental Creek, in that area, there is an aggregate thickness of well over 1,000 feet of silicic tuff, flow breccia, lava, and conglomerate, probably including mud flows, plus some sandstone and shale (Ross 1933a, p. 590-592; Shenon, Ross 1936, p. 10-11). Some of the conglomerate high in the sequence contains well-rounded pebbles of rhyolitic rocks 1-2 inches in diameter. These rocks are all somewhat kaolinized, at least in the vicinity of mines, where most available data were obtained. The area also contains conglomerate at the base of the Challis Volcanics in several places. The conglomerate contains poorly rounded and assorted pebbles and boulders of pre-Tertiary rocks, in a matrix that includes volcanic materials.

The Challis Volcanics in the Hailey quadrangle (Umpleby, Westgate, Ross 1930, p. 35-36) include fairly large amounts of clastic rocks. Much of this is tuffaceous, in part the direct product of volcanic eruptions, in part water-laid and admixed with sedimentary material. The tuff is cream-white, light brown, gray and green, but green and brown colors predominate. The rock contains crystal fragments in a fine-grained, in part, glassy groundmass. Some of it corresponds in composition to latite and rhyolite but others correspond to augite andesite and contain bombs of that lava. There are all gradations between tuff and sandstone, conglomerate and shale, of fluviatile origin.

Tuffaceous beds interbedded with lava are conspicuous near Pass Creek in the Lost River Range. Locally they rest on Paleozoic rocks. They are reported (Umpleby 1917, p. 35) to have an aggregate thickness of about 2,000 feet. In the Mackay quadrangle, tuff is present at various horizons but nowhere in mappable quantities. The quadrangle, however, has conglomerate at the base of the Challis Volcanics in numerous places.

Conglomerate more or less like that conspicuous in the Mackay quadrangle is probably present at or close to the base of the Challis Volcanics in numerous areas. It is in individually small bodies, mostly deposited in old stream valleys. Where the volcanics are still plentiful such deposits would be largely hidden from view. The Challis Volcanics rest on an irregular surface with original gradients steeper in some places than in others (Ross 1961b) so the distribution would be correspondingly variable.

In descriptions already given conglomerate at or close to the local base of the formation has been mentioned in and near the Germer Basin and at the uranium prospects near Stanley.

Anderson (1959, p. 21-22; 1961,p. 30-31) reports beds in which conglomerate is plentiful below the flows of the Challis Volcanics in the North Fork and Lemhi quadrangles. In the North Fork area he called these beds the Kriley Formation. In the Lemhi area he termed them the Beaverhead Conglomerate on the assumption that they were stratigraphically equivalent to the Beaverhead Formation in Montana (Lowell, Klepper 1953). The most probable assumption from his descriptions is that his Kriley Formation and his Beaverhead Conglomerate are both among the coarse stream deposits laid down at the beginning of the crustal disturbances related to the Challis volcanism.

Tuffaceous beds analogous to the Germer Tuffaceous Member are plentiful near Salmon (Anderson 1956, p. 22-26). In that area the tuffaceous unit includes all the volcanic rocks above the latite-andesite member and is particularly well displayed in T 20 N, R 21 E, although it extends east and north of that township. The dominant component is well-bedded, light-colored tuff but includes lava flows and welded tuff that have not been mapped separately.

The tuff is fairly uniform in character throughout. Most of it, especially low in the sequence, is conspicuously stratified in beds less than an inch to a foot or more thick. The rock is mostly light gray and of medium grain. Some beds high in the sequence are fairly coarse. Much of it is the product of ash falls but some has been sorted in bodies of shallow water. Some

beds are shaly. The tuff contains quartz, plagioclase, potassium feldspar, biotite and bits of volcanic rock in a matrix composed largely of partly devitrified ash. Some shards are discernible. At the north end of the Lemhi Range there are deposits of breccia in which small fragments of volcanic rock and of quartzite are firmly held in a matrix of partly devitrified glass.

The principal flows intercalated in the tuffaceous unit of the area consist of red, pink and lavender porphyritic quartz latite or rhyolite, topped by black obsidian. The constituents include quartz, sanidine, feldspar (largely oligoclase or sodic andesine), biotite and, locally, basaltic hornblende with minor quantities of zircon, magnetite and apatite. In some flows the groundmass is microspherulitic, in others it is partly devitrified glass.

Some of the welded tuff has colors similar to those of the flows but most of it is white or light gray. Conspicuous phenocrysts of quartz, plagioclase and biotite are embedded in a groundmass composed mainly of welded, partly devitrified ash and shards, with patches of material composed of tiny microspherulites mingled with partly devitrified glass and streaks of pumice. Most of the welded tuff contains numerous small fragments of volcanic rock.

Anderson (1956, p. 23) estimated the thickness of the dominantly tuffaceous unit as 4,000 feet or more.

As noted above, the valleys of the Salmon and Lemhi Rivers near Salmon contain, in addition to the dominantly volcanic rocks already described, substantial thickness of beds that are largely sedimentary. Close to Salmon, Anderson (1956, p. 28-31) originally mapped the dominantly sedimentary beds as constituting his Carmen Formation, which he later (1959, p. 27) renamed the Kirtley Formation on the assumption that the name Carmen is preoccupied. In the area around Salmon, Anderson recognized the presence of thousands of feet of tuffaceous beds but did not distinguish them in mapping. Apparently the Carmen Formation of his original usage is regarded by him as deposited unconformably on both the Precambrian rocks and the Challis Volcanics, including the tuffaceous beds in the latter. Near Salmon, it may include equivalents of most of the Tertiary sedimentary units he mapped separately in neighboring areas.

He notes that some of the tuffaceous beds belonging to the Challis Volcanics may have been included in the Carmen Formation on his map. He says the unit consists of moderately well-indurated, thin-bedded, fine-grained clastic rocks. Shale predominates but there is also much sandstone and, especially near the base, conspicuous conglomerate. The shale contains bits of quartz,

feldspar, mica, ash, and clay. Some of it has numerous plant fragments and thin beds of lignite are included. Some of the shale is bentonitic. Some of the lignite was formerly mined for local use. All the plant fossils found near Salmon came from more or less lignitic beds in the Carmen Formation. Some of the conglomerate rests, apparently unconformably, on tuff.

Anderson believes that his Carmen Formation contains no lava, welded tuff, or other product of direct volcanism. However, Behre's (1930, p. 240-243) detailed study at a quarry for building stone in the Carmen Formation shows that tuff is included. As most of the tuffaceous rocks of the Challis Volcanics are lithic and crystal tuffs, more or less water-sorted, it would be difficult to distinguish between these and the sandstone and shale in the Carmen Formation supposedly derived from tuff by erosion.

Anderson reports an unconformity between his Carmen Formation and flows of the Challis Volcanics but the evidence adduced on this point is not convincing. A change from the eruption of lava to the deposition of tuffaceous sediments implies interruption in the depositional sequence but not necessarily a significant time gap. Similarly unconformities are reported between the different sedimentary units that Anderson has distinguished in the valleys of the Salmon and Lemhi rivers but the significance of these is difficult to evaluate. They must record interruptions in sedimentation.

Anderson (1959, p. 27) states: 'As originally applied in the Salmon quadrangle, the name Carmen was given to all the Tertiary sedimentary strata within the Salmon Basin (Anderson 1957, p. 28-30); but when study of the Baker quadrangle was begun, the Tertiary beds were found to contain three unconformable formations, for one of which the name Carmen was retained (Anderson 1957, p. 15-19). Later, when it was learned that Carmen had been used as a formational name elsewhere, the name Kirtley was accepted and applied to the youngest of the Tertiary Formations throughout the Salmon Basin.' In areas outside the Salmon quadrangle Anderson (1957, p. 16-19; 1959, p. 24-28; 1961, p. 29-35) described the Kenney and Geertson Formations as underlying the Kirtley Formation. The statement quoted above, coupled with Anderson's geologic maps indicates that the Carmen Formation as originally mapped includes the Kenney and Geertson Formations. Further, comparison of Anderson's maps indicates that his Kenney and Geertson Formations in the northern part of the Lemhi quadrangle are included in the Challis Volcanics in the southern part of the Baker quadrangle. The three dominantly sedimentary units that Anderson calls the Kenney, Geertson, and Kirtley Formations are regarded in the present report as components of the Challis Volcanics. Reasons

for this decision are brought out in the descriptions below and in the discussion of available data in regard to age that follows.

Anderson (1961, p. 31-32) says that in the Lemhi quadrangle his Kenney Formation is exposed on the lower slopes flanking the valley of the Lamhi River and near the base of the Beaverhead Mountains. The unit is dominantly sedimentary but includes much tuffaceous material, including some welded tuff, especially in its lower part. Most beds are light colored, almost white. Locally, near the base, the formation has conglomerate composed of pebbles and boulders of volcanic rocks. The proportion of shale and sandstone increases upward and these rocks predominate in the upper part of the unit. Some of the shale is bentonitic.

The Kenney Formation of Anderson (1958, p. 16-17) is exposed in hills on both sides of the Lemhi River, especially on the northeast. It is white to pale buff and composed largely of sedimentary materials, derived in part by erosion of lava of the Challis Volcanics. Some tuff and bentonite are included. Locally the base is coarse conglomerate or breccia composed of material of andesitic and latitic composition. In most places shale predominates and sandstone, conglomerate, tuff and bentonite are subordinate and not individually persistent. Much of the sandstone is composed of material of igneous derivation.

The Kenney Formation is exposed in the southeastern part of the North Fork quadrangle (Anderson 1959, p. 25-26). The unit contains much the same sorts of rocks as in the area along the Lemhi River near Baker, described in the preceding paragraph, but conglomerate is much more abundant in it. The components of the conglomerate in the North Fork quadrangle are derived mainly from rocks of the Belt Series, in contrast to those near Baker, which came mainly from the Challis Volcanics.

Anderson (1961, p. 33-34) maps his Geertsen Formation as overlying his Kenney Formation in the Lemhi quadrangle where it is exposed only on the east side of the river valley. The Geertson is light colored and contains shale, sandy shale, sandstone, conglomerate and thin local beds of bentonite or bentonitic shale. In the Lemhi quadrangle the proportion of shaly beds increases to the south.

In the area around Baker (Anderson 1957, p. 17-18) the Geertson Formation is similar to beds so designated in the Lemhi quadrangle, except that thin beds of impure lignite are recorded. To the north, in the North Fork quadrangle (Anderson 1959, p. 26-27) the Geertson Formation is the most widespread of the Tertiary sedimentary rocks. It is described as differing from the Geertson in areas described above. In the North Fork area the

beds that have been mapped as Geertson are described as mainly conglomerate with the characteristics of fanglomerate. There are thin partings of sandy shale and southward within the map area the proportion of shale increases.

The Kirtley Formation, which is a renamed and restricted equivalent of the Carmen Formation, first described near Salmon (Anderson 1956, p. 28-31) is plentiful in the Lemhi quadrangle, on both sides of the valley of the Lemhi River. It consists largely of white, gray and tan thin-bedded, fine-grained shale with some sandy beds and thin bentonite layers. Some of the shale is highly siliceous, some is sandy or silty and some contains vegetable debris. Locally some beds contain pebbles. When the area around Baker was mapped (Anderson 1957, p. 18-19) the name Carmen Formation was applied to beds later called the Kirtley Formation. These are essentially identical to those of the Kirtley Formation in the Lemhi quadrangle, just described.

When the area around North Fork was mapped (Anderson 1959, p. 27-28) the upper part of the old Carmen Formation was called the Kirtley Formation. Only a little of this formation extends north from the Salmon quadrangle into the North Fork quadrangle. There it contains well-stratified shale similar to that in the Baker and Lemhi areas, with a little conglomerate at or near its base. No bentonitic or lignitic beds were noted."

Yankee Fork Rhyolite

Distribution. "The Yankee Fork Rhyolite takes its name from the Yankee Fork of the Salmon River and consists of rhyolitic material of distinctive appearance high in the Challis Volcanics and commonly interbedded with tuff. The member has been mapped so far only in and close to the Casto and Bayhorse quadrangles. In the former it is prominent around the head of the Yankee Fork and near Meyers Cove, with scattered exposures in other parts of the quadrangle. It appears to be abundant east of the Casto quadrangle in areas not yet mapped in detail. In and close to the Bayhorse quadrangle, the Yankee Fork Rhyolite is broken up into several units by interleaved tuffaceous beds and three varieties of the rhyolite are distinguished. The principal exposures, all small, are on the borders of Round Valley near Challis and in Ts. 9-11 N, Rs. 17-19 E."

Character. "Certainly much of the rock mapped as Yankee Fork Rhyolite is rhyolitic welded tuff rather than lava. Some may well be glassy lava. The distinction was not in mind in 1924 to 1930 when the fieldwork in the Casto and Bayhorse quadrangles was done. In 1953 some of the

exposures of the Yankee Fork Rhyolite in the Bayhorse quadrangle were revisited in company with Howard Powers. This visit, followed by petrographic work by Powers, shows that, while some of the rock is of indeterminate origin, much of it is welded tuff. It may be remarked that at the time of the original investigation certain beds in the Germer Tuffaceous Member (Ross 1937 (1938), p. 54-55) were recognized as composed of glassy pumice fragments that had welded together during flowage. That is, these beds are welded tuff.

Most of the rocks mapped in the Casto quadrangle as Yankee Fork Rhyolite (Ross 1934a, p. 52-53) are brownish rocks with a more or less distinct violet tinge. Some, especially where weathered, are light colored, but a few are almost black. Most of them are characterized by round, dull purplish black phenocrysts of quartz about 1.25 mm in average diameter. The groundmass was largely originally glassy but contains quartz, orthoclase and sodic plagioclase. Some have distinct flow structure, but most do not. Many of the lower flows contain subangular fragments of the light-colored rocks below them, rarely as much as an inch in maximum dimension. These fragments, coupled with local irregularities in the lower contacts, indicate that an erosion interval, probably short, preceded its eruption. In most localities light-colored, clastic tuff is interbedded with the rhyolite but everywhere in decidedly minor amount. In hundreds of feet of rhyolite there may be from one to half a dozen such beds from a few feet to a few score feet in aggregate thickness.

The most abundant and widespread variety of the Yankee Fork Rhyolite in and near the Bayhorse quadrangle (Ross 1937 (1938), p. 59-63) is characterized by abundant smoky quartz phenocrysts and a groundmass of stony rather than glassy appearance. It is closely similar to the majority of the beds composing the member in the Casto quadrangle, described above. This variety is well exposed in the bluffs immediately north of Challis, and at that locality, is welded tuff. The rhyolite in bluffs across the river from Challis is probably also welded tuff, although samples from there are so devitrified as to abscure the original texture. Another variety also has a groundmass of stony or lithoidal appearance but tends to be lighter colored and locally more brecciated. The quartz phenocrysts are less smoky and also less abundant and feldspar phenocrysts are more conspicuous than in the first variety mentioned. This variety may be lava rather than welded tuff. The third phase distinguished in the Bayhorse quadrangle differs considerably in appearance from the other two but is much like many of the darker beds included in the Yankee Fork Rhyolite in the Casto quadrangle, especially near Meyers Cove. Its groundmass is nearly black and has a

distinctly glassy appearance. The phenocrysts are white feldspar (largely oligoclase) and quartz; locally the rock is mottled with areas of pink feldspar. A little biotite is present. A sample of this third variety was found by Powers to be welded tuff, but other samples retain no characteristics indicative of such an origin.

The chemical composition of components of the Yankee Fork Rhyolite is indicated by the 3 analyses below. Analyses of sample I. C. 269, from the type locality of the member, and of sample 53-P-73, from the most abundant variety in the Bayhorse region, resemble each other. The partial analysis of M27 from another variety in the Bayhorse region, corresponds to a somewhat more calcic rock than the other two. All 3 samples are silicic rocks, but M27 verges on a latitic composition.

Three of the samples of Yankee Fork Rhyolite collected by Powers have received spectrographic analysis. The results are listed here so as to be available for anyone interested. All are from the Bayhorse region. A standard analysis of sample 53-P-73 for major constituents is given in Table 5. The partial analysis of sample M27 given in the same table came from the same exposure as sample 53-P-77.

Wherever the Yankee Fork Rhyolite is exposed in the Casto quadrangle it is much eroded. In the vicinity of Yankee Fork less than 1,000 feet of it remains, although it is probable that the maximum original thickness was considerably greater. West of Camas Creek the thickness is probably much less than 1,000 feet. In exposures about two and a half miles south of Meyers Cove and a short distance east of the Casto quadrangle 1,600 feet of rhyolitic rock with a few score feet of interbedded tuff are present. To the southeast the thickness decreases greatly. On upper Mill Creek, just outside of the southeast corner of the Casto quadrangle, probably little over 300 feet are present. Immediately north of Challis, the thickness is also somewhat over 300 feet. In the areas farther south in the Bayhorse quadrangle, the flows in some exposures are only about 25 feet thick and the thickest mass, that in T 11 N, R 19 E, is little more than 100 feet thick." (Ross' Challis description.)

Subsection 1 - Bellevue Foothills Subsection

Subsection 1 covers the foothills of the Smokey Mountains and the lower part of the drainage of the Little Wood River including Muldoon Canyon. As a foothills area it is very dissected with low to moderate relief as compared to the rugged mountain terrain to the north.

Anderson and Wagner (1946) discussed the geology of the Smokey Mountains. Basalts of Challis age or Paleozoics (mostly Wood

River with a few spots of Milligen) overlie Idaho batholith granitic rocks. Anderson noted that several Mesozoic age fault systems in the batholith controlled aplite and pegmatite dike intrusives. Dikes and faults trend N 10° to 20° E. Anderson also mentioned a set of Tertiary (?) quartz veins and faults which strike N 50° - 70° W. Examination of the Hailey linear map shows both sets of faults. The high concentration of faults on this region reflects the large number of mines (Hailey - Bellevue Gold Belt).

The eastern part of the Subsection is capped by Challis volcanics (latite-andesite member, Tee2) or the Wood River Formation. The Wood River Valley is filled with Quaternary alluvium.

Subsection 2 - Ketchum Subsection

Subsection 2, like Subsection 1, is covered by Wood River (Pm) and Milligen Formations (Mu) or Challis volcanics (Tee2). Idaho batholith outcrops along the western border of Subsection 2. Anderson, Kiilsgaard, and Fryklund (1950) studied the Mineral Hill and Warmsprings Mining Districts which are in the southeastern part of the Subsection.

The Paleozoics in the mining districts are folded into a broad, broken anticline the axis of which parallels the Wood River Valley. Many minor, complicated drag folds occur in the Milligen Formation. The emplacement of the Idaho batholith further complicated the structure and may have caused some of the thrust faults in the area.

Extrusion of the Snake River Plain basalts in Tertiary time was followed by subsidence of the Snake River Plain and uplift of the area north of the plain. The broad aggraded floor of the Wood River Valley, which is filled with alluvium, may have been dammed up to the south by the Snake River flows. Anderson, Kiilsgaard, and Fryklund described four sets of terrace gravels in the valley and in higher elevation drainages which are in part due to several periods of Pleistocene glaciation.

The northern half of Subsection 2 is covered by Challis volcanics. The Hailey linear map shows that there are fewer faults (?) in the Challis volcanics than in the Paleozoics. The linears in the Challis have a strong northeast trend while faults in the Paleozoic trend N 70° - 80° W and N 10° - 20° E (Anderson, Kiilsgaard, Fryklund). This probably means that the northwest faulting is older than the northeast faulting.

Subsection 3 - Boulder-Pioneer Subsection

Subsection 3 is conveniently divided into three regions. The eastern section includes the part of Custer County south of Summit Creek which is the Pioneer Mountains (the Pioneers include all of the area labeled mpC_SHy or mpC_SE for the Hyndman Peak and East Fork Formations). The southwestern part of the Subsection includes the large thrust belt from Ketchum to the southern edge of the Subsection. The northern part of Subsection 3 is covered with Challis volcanics.

The northern part of 3 is based on the geologic map of the Bayhorse Region (Ross, 1937) while the Hailey 15 minute quadrangle is

based on Umpleby, Westgate, and Ross' geologic map (1930). When the Bayhorse Region is compared with the Hailey quadrangle there is a considerable difference. Rock types agree fairly well but there is substantially more structure shown on the Hailey map than the Bayhorse map. The Bayhorse area undoubtably has as much structure as the Hailey area so structurally the maps are a bit misleading.

The Pioneer Mountains section divides the Wood River drainage from the Big Lost River drainage to the east. It is an area in which much of the geology of central Idaho comes together. It is the only part of Subsection 3 which contains possible Precambrian metasediments. On the western side of the Pioneers the geology consists of thrust and folded Paleozoics (Wood River and Milligen Formation primarily) while east of the mountains the Paleozoics have different lithologies (Copper Basin Group and White Knob limestone). The central part of the Pioneers may be near the eastern border of the batholith.

Umpleby, et. al., (1930) described two units of metasedimentary rocks which were named the Hyndman Peak Formation (mpC_sH_y) and the East Fork Formation (mpC_sE). These units are found at the head of Hyndman Creek and the East Fork of the Wood River which are located east of Ketchum on the Hailey AMS map. Dover (1969) updated the original work and the following rock descriptions are his.

The Hyndman Peak Formation (about 5,200 feet thick) is divided into four members which are labeled A (base) to D (top). Member A is a pelitic schist (200 feet thick). Member B is a gneissic quartizite (1,200 feet thick). Member C, a green banded calc-silicate rock is 600 feet thick and Member D is another gneissic quartzite (1,400 feet thick).

The East Fork Formation (about 4,000 feet thick) is also divided into four members. Member E is a 700 feet thick dolomitic marble, Member F is a 300 feet thick vitreous quartzite, Member G is an 800 feet thick marble and Member H is 2,000 feet of what may be Kinnikinic quartzite.

Dover suggested that these units may be Precambrian Belt Supergroup equivalents. Almost all of the metasedimentary rocks in the Rocky Mountain Province are believed to be Belt equivalents as will be pointed out in later discussions. The Hyndman Peak and East Fork Formations are overlain by typical Paleozoic sediments.

Beneath the two Precambrian formations is a sheet like intrusion of Idaho batholith and this sheet is underlain by a migmatite complex which is exposed in the central part of the Pioneers at the heads of Kane Creek and upper Wildhorse Creek.

It should be noted that the sequence, thrusted metasediments overlying batholith with batholith overlying migmatites is essentially the model proposed by Hamilton (1973) for the Boulder batholith in Montana. Dover said that the batholith was synorogenic but described several small quartz monzonite intrusions in the area which he called post orogenic. These post orogenic intrusives appear to be similar to the White Cloud Stock (Bennett, 1973) and the leucocratic

intrusives in the Sawtooths (Reid, 1963). These may all represent late stage Idaho batholith differentiates. Dover's description of the granitic rocks associated with the Idaho batholith intrusives and the post orogenic intrusives are included here.

Synorogenic Cretaceous Intrusive Rocks of the Pioneer Mountains Pluton

Introductory Statement

"Gneissose and granitic rocks occupy 75 square miles in the heart of the Pioneer Mountains. The partly gneissose plutonic rocks that nearly surround the migmatites are to be considered here."

Areal Distribution

"The plutonic rocks form an intrusive sheet that separates the Wildhorse Canyon migmatite complex from overlying metasedimentary rocks. The pluton has an arcuate outcrop belt that occupies about 50 square miles. The belt widens from one mile at the apex of the arc near the head of the East Fork of the Big Wood River to six miles on the northeast flank of the arc in the Broad Canyon area. The pluton is well exposed along the crest of the Pioneer Mountains and in the rugged interior of the northeastern part of the range."

General Description

"The granitic rocks range in composition from clinopyroxene- hornblende-biotite quartz diorite and mafic granodiorite to leucocratic hornblende-biotite quartz monzonite. The mineral constituents and paragenetic sequence of these various rocks are identical except for the restriction of minor clinopyroxene to the quartz dioritic and mafic granodioritic rocks. Small bodies approaching pyroxenite in composition are locally associated with the quartz diorite.

The rocks are coarse-grained and range from even-grained and moderately strongly gneissose to directionless and coarsely porphyritic. The gneissose structure is marked by aligned mafic minerals. K-feldspar and quartz constitute the megacrysts of the coarsely porphyritic phase.

Although the plutonic rocks vary considerably in composition and texture from place to place, these variations are gradational and the rock is quite homogeneous at any one locality. The subtle gradations occur over distances of several miles. Sampling during the present study permits delineation of the major compositional varieties only within very wide limits. However, there is a general transition from west to east across the outcrop

belt from the gneissose quartz diorite - mafic granodiorite phase to the directionless quartz monzonite phase. Deviations from this trend occur locally. All of these rocks are considered to have been emplaced during a single intrusive event."

Field Relations

"Field evidence establishes the intrusive character of the granitic rocks. They show clear cross-cutting relations with the two adjacent metamorphic units. Fault contacts separate the plutonic rocks from all other map units.

The intrusive contact with the underlying Wildhorse Canyon migmatites is irregular but generally sharp. Little or no reaction has occurred along this contact, probably owing to the low reactivity of the country rocks.

The contact with the overlying metasedimentary rocks is much more variable. The contact with gneissose quartzite members is sharp; only small, irregular granitic apophyses project into the quartzite. Near the main intrusive mass, the banded calc-silicate member is characterized at most places by thin- fine-grained granitic layers parallel to the banding. The contact with the pelitic schist member is marked by abundant sills and lenses of intrusive material and by a wide assimilation zone containing lenticular schist inclusions and schlieren in all stages of digestion. Some andalusite schist inclusions contain late fibrolite that probably reflects a pressure increase rather than a temperature rise. The pressure increase may mark a time when the pluton achieved a degree of consolidation sufficient to transmit shearing stresses. Huskier sillimanite needles also occur in the feldspar of intrusive material within a few centimeters of the schist inclusions. This sillimanite crystallized under metamorphic conditions either from alumina released from pelitic schist assimilated in situ, or from alumina that migrated from adjacent inclusions. A variety of hybrid rocks and feldspathized schistose cognate inclusions also occurs in the assimilation zone.

> Assimilation zone. In the assimilation zone occur all gradations between (1) fresh and alusite-muscovite-biotite schist with incipient retrogression of andalusite to muscovite and biotite to chlorite, (2) feldspathized schist with more advanced andalusite alteration and incipient segregation of biotite, and (3) hybrid igneous rocks enriched in quartz and having somewhat higher color index than usual. The hybrid rocks contain small biotite schlieren up to a few cms.

long which are "basified" schist remnants; a few muscovite pseudomorphs after andalusite are also preserved. Schist-derived tourmaline occurs in the hybrid rocks. Faint "ghosts" of nearly digested leucocratic schist lenses can be detected in the pluton near the assimilation zone.

Fibrolite stringers clearly cross-cut and therefore postdate poikioblastic andalusite in some schist inclusions; in other samples retrogressive muscovite around andalusite obscures the fibrolite relationship. Sillimanite occurs in adjacent intrusive rocks as radiating needles and as needles aligned along crystallographic directions in plagioclase.

Schistose leuco-biotite amphibolite occurs locally in the assimilation zone. Lenticular clusters of small hornblende grains aligned along the schistosity may represent breakdown of larger crystals. The amphibolite contains large feldspar porphyroblasts with identical composition (An30 to An40), euhedral oscillatory zoning and blocky growth twinning as adjacent igneous feldspar. Some samples contain 5-10 per cent of plagioclase porphyroblasts. Irregular zones of amphibolite with lower color index give the rock a mottled appearance. The amphibolite is probably an early border phase of the pluton that underwent extensive metamorphic recrystallization and later incipient felspathization related to intrusions."

Post-Orogenic Quartz Monzonite Intrusions

Areal Distribution

"Three relatively small intrusions of quartz monzonite crop out near the crest of the Pioneer Mountains. The largest and best exposed of the bodies occupies about three square miles along the valley of Summit Creek. Smaller isolated offshoots from the main Summit Creek intrusion occur in the canyon of Little Falls Creek and on the ridge northeast of the Big Falls Creek. Another poorly exposed intrusion occurs eight miles to the southeast on the divide between the two main forks of Hyndman Creek. The third intrusion occupies a small cirque on the west side of East Fork Canyon."

Field Relations

"Exposed contacts of the intrusive bodies cut across country structures. The Summit Creek pluton made room

for itself by doming up the Ordovician country rocks; post-orogenic intrusion probably was accompanied by uplift throughout the range. An offshoot of the Summit Creek pluton is injected across a major thrust fault into coarsely clastic rocks of the northeastern Pioneer Mountains succession. The Hyndman Creek and East Fork intrusions are in medium- to high-grade metasediments. Surface exposures of the Summit Creek and Hyndman Creek bodies are altered and crumbly; the East Fork intrusion is fresh."

Petrographic Description

"The Summit Creek and Hyndman Creek intrusions consist of directionless, medium- to coarse-grained biotite-quartz monzonite. The East Fork intrusion is similar in composition but is finer grained and very weakly gneissose. Typical samples from all of these bodies consist of 25 to 35 per cent quartz, 25 to 35 per cent microcline, 30 per cent to 40 per cent plagioclase, and 5 per cent or less biotite. Some samples contain sieved remnants of biotitized ferrohastingsitic hornblende. Plagioclase is subhedral and shows blocky primary growth twinning, synneusis texture and simple subhedral to patchy normal zoning from AN30 to AN20. Relatively calcic cores are extensively altered to sericite and minor pistacite in some sections. Late-magmatic quartz and perthitic anorthoclase megacrysts appear to have formed by replacement from deuteric fluids enriched in sodium. Advanced alteration to kaolinite? imparts a dusty appearance to the anorthoclase. Magnetite, apatite, allanite, muscovite and euhedral sphene occur in traces.

The quartz monzonite is entirely igneous in character. Minor crystal bending, weak undulatory extinction, and occasional broken crystals reflect final adjustment in the nearly consolidated bodies.

A comparison with the syn-orogenic Pioneer Mountains pluton indicates that the post-orogenic quartz monzonite contains (1) a more sodic plagioclase, (2) anorthoclase, (3) only 5 per cent biotite, (4) less abundant heavy minerals, and (5) some accessory muscovite. Only the East Fork intrusion displays evidence of cataclasis. Its weak gneissose structure may result from emplacement during a time intermediate between that of the syn-orogenic Pioneer Mountains pluton and the directionless post-orogenic intrusions." (Dover, pg. 37, 38 and 39)

The southwestern part of Subsection 3 is similar to parts of Subsection 2 except that more of the area is covered by Milligen Formation than Wood River Formation. Kiilsgaard (Anderson, Kiilsgaard, Fryklund) 1950, suggested that the major drainages

which flow into the Wood River are fault controlled. The Hailey linear map shows a strong tendency for the streams to coincide with linears supporting Kiilsgaard's observations.

The Milligen Formation is the main structural agent in the area and Kiilsgaard noted that the Milligen Formation was often separated from the overlying Wood River Formation by thrust faulting. The trend of these thrust faults on the Hailey quad map is to the northwest. Kiilsgaard suggested that thrusting occurred in the Laramide Orogeny (Late Cretaceous - Early Tertiary). Minerialization in the area is controlled by strong, northwest trending, shears in the Milligen carbonaceous argillites.

The northern part of Subsection 3 contains the Boulder Mountains (western edge of the range runs from Galena Summit along the Big Wood River to the North Fork of the Wood River near Boulder Peak). The Boulder Mountains are poorly mapped but are much more complex than shown by Ross (1937). The Boulder Mountains including Boulder Peak, Glassford Peak, and Ryan Peak form a major divide between the East Fork of the Salmon River which flows north, and the Big Lost River drainage which flows south.

The western face of the Boulders (drained by Cherry Creek, Spring Creek, King Creek, and Boulder Creek) is composed of granitic rock of the batholith. An Aerial Magnetic Survey flown by the U.S. Geological Survey (White Clouds Primitive Area Study) shows that the Boulder granitic mass may be linked to the White Cloud stock to the north (Challis, AMS map). The Boulder granitics are capped by complexly thrust faulted Milligen and Wood River units. The Paleozoics are in turn capped with Challis volcanics. Paleozoic units outcrop in the volcanics in several areas of the East Fork of the Salmon. All of this new data will be included in the White Clouds Primitive Area Study by Charles M. Tschanz (U.S.G.S., Denver) which will be released in 1974.

Most of the area north of the Boulders is covered by Challis volcanics (mostly the latite-andesite member Tee2) but the volcanics are partially intruded by hundreds of Tertiary dikes and sills. Many of these dikes are porphyries and sometimes form small stocks. Tertiary dikes are mentioned frequently in the State literature and while they are described in detail little mention is ever given to the number of these intrusives. Taken in total the dike complexes must represent a considerable amount of magma.

Subsection 4 - White Clouds Subsection

Subsection 4 is located in the southeast corner of the Challis AMS map. The major feature of the subsection is the White Cloud stock which is surrounded by the Milligen and Wood River Formation. The stock is mostly quartz monzonite (core slightly more felsic than the outer edges). Bennett (1973) theorized on the basis of geochemistry and petrology that the stock represented a late phase of the Idaho batholith which was volatile saturated and because of this forcibly emplaced. There are extensive apophysis and fractures

in the roof rocks of the stock and all of the roof Paleozoics have been contact metamorphosed (the White Clouds are white because the Paleozoics are recrystallized to calc silicates). Evidence shows that the stock magma flowed out from under a cover of solidified, passively emplaced, low volatile, early stage, Idaho batholith.

The area surrounding the stock was mapped by Ross (1937) and he noted that the three to four mile strip of Wood River rocks which is located between the stock and a pie shaped section of batholith directly west of the stock was folded into a northeast trending anticlinal structure. The White Cloud stock itself is bordered on the east by a major fault (the Uncle Jess fault) and on the west by a major fault which is associated with the anticlinal fold. Most of Subsection 4 has been alpine glaciated. The rocks in the area are only slightly weathered because high elevations (about 9,000 feet) favored Alpine glacial scouring.

Subsection 5 - East Fork Salmon River Subsection

Subsection 5 covers a large part of the Bayhorse Area. Major drainages in the Subsection include the Salmon River, and the East Fork of the Salmon River mapped by Ross (1937). The area is covered with volcanics with good exposures of Kinnikinic quartzite (Om) and Laketown dolomite (Sm).

The major structure in the region is an anticline near Bayhorse and Clayton. This is one of the major structures in this part of Idaho. Ross believed that much of the deformation in this region was associated with the intrusives of the Idaho batholith. Later, milder deformation occurred in Tertiary times when the Challis volcanics were slightly deformed. Ross suggested that the northeast trending fault system (see Challis linear map) is Tertiary in age whereas the northwest system is probably Mesozoic (batholith associated).

Ross's descussion of the Challis volcanics explains much about the present topography. The volcanics were extruded into an area of considerable relief (probably similar to the present topography). This topography resulted in much slumping and internal deformation by gravity sliding of flows and ash falls. Figure 17 shows Ross's restoration of the Challis surface after deformation but without erosion. Ross said that the entire Bayhorse area was covered by volcanics except for the very highest peaks. Variations in deposition of the volcanic materials are also evident in Figure 17. Low hills in and near Section 13, Township 12N, Range 20E, east of Grand View Canyon (see Figure 16 for reference) are covered by Yankee Fork rhyolite which rests directly on Jefferson dolomite without any latite-andesite or Germer tuffaceous members. The Post Challis surface was mature and is reflected in many sections of the Bayhorse Region by flat ridge crests which are near or above 9,000 feet. Railroad Ridge, just north of the White Cloud stock is a remnant of this surface.

Ross noted that two stages of glaciation had affected the region. The Nebraskan glacial period is evidenced by glacial material which

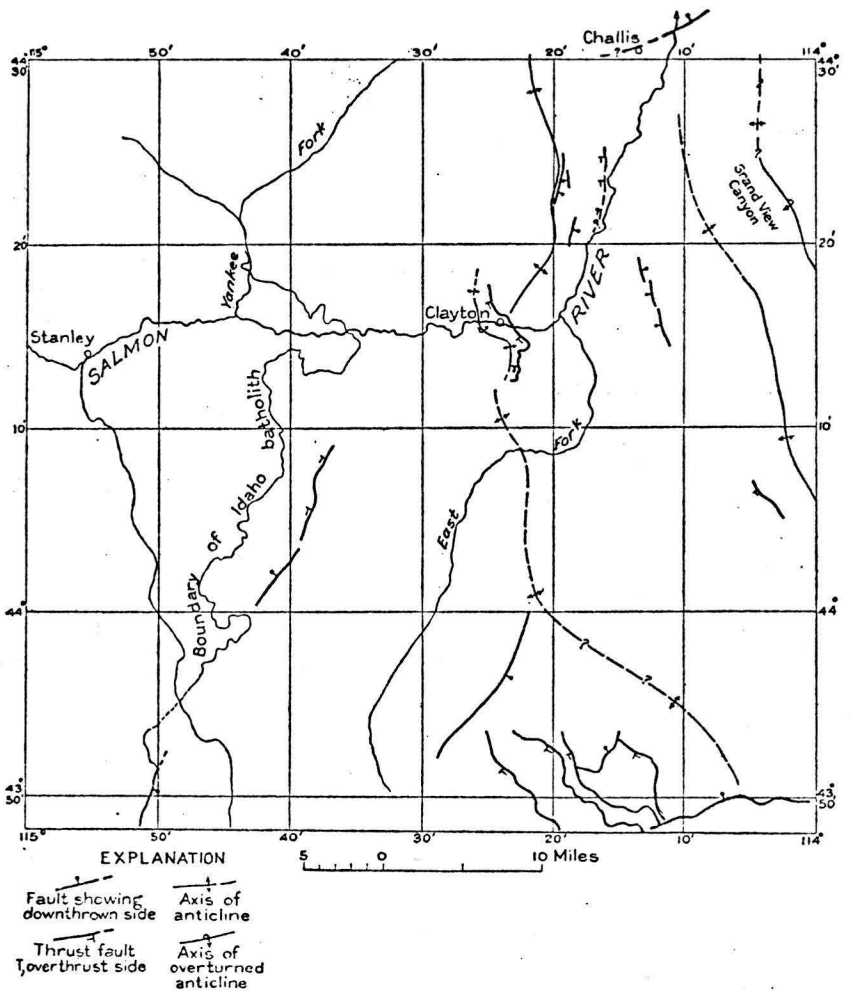


Figure 16. Sketch showing principal structures in the Bayhorse area (Ross, 1937).

has itself been glaciated during the latter Wisconsin glacial period. Some of this material covers the top of Railroad Ridge. Figure 18 shows a restoration of the surface on which some of the early Pleistocene glacial material was deposited. Nebraskan glacial deposits are found only above 9,000 feet. Complex erosion followed the Nebraskan period as is evidenced by several sets of terrace gravels at different elevations. This dissected surface was glaciated in Wisconsin times and evidence of the glaciation is found at elevations ranging from 6,500 to 8,000 feet. The higher ranges like the White Clouds were intensly eroded by these glaciers. The present topography is similar to that of Wisconsin time with only slight, subsequent fluvial erosion.

Subsection 6 - Challis Subsection

Subsection 6 contains the area which is drained by the Salmon River from Challis to Salmon on the northwest side of the river. Ross in 1963a described a road log along U.S. Highway 93 from Challis to Gibbons Pass in Montana. The Challis AMS sheet contains the part of the log from Challis to the Pahsimeroi River and then along the Salmon River onto the Dubois AMS map.

The major unit in the southwestern part of 6 and along the Salmon River to Rattlesnake Creek is Challis volcanics (undifferentiated for the most part) with a few outcrops of Kinnikinic quartzite (Om). Just to the northwest of Rattlesnake Creek and on a line to Iron Lake is the contact between the Challis volcanics and undifferentiated Precambrian metasediments (Swauger and Lemhi quartzites, mpCsud). The Challis linear map shows that a long northwesterly trending linear coincides with Morgan Creek.

Subsection 7 - Salmon Mountain Subsection

Subsection 7 includes the Salmon River Mountains which form the western border of the Salmon River in this region. Seven has been mapped by Shockey (1957, western part) and Anderson (1956, eastern part). Rock units in the subsection consist of undifferentiated quartzites (Swauger/Lemhi, mpCsud) and areas mapped as Lemhi quartzite (mpC_s2). Some Idaho batholith rocks and Challis volcanics occur in the region also. The general structure in the area is discussed in Section IX, Subsection 4.

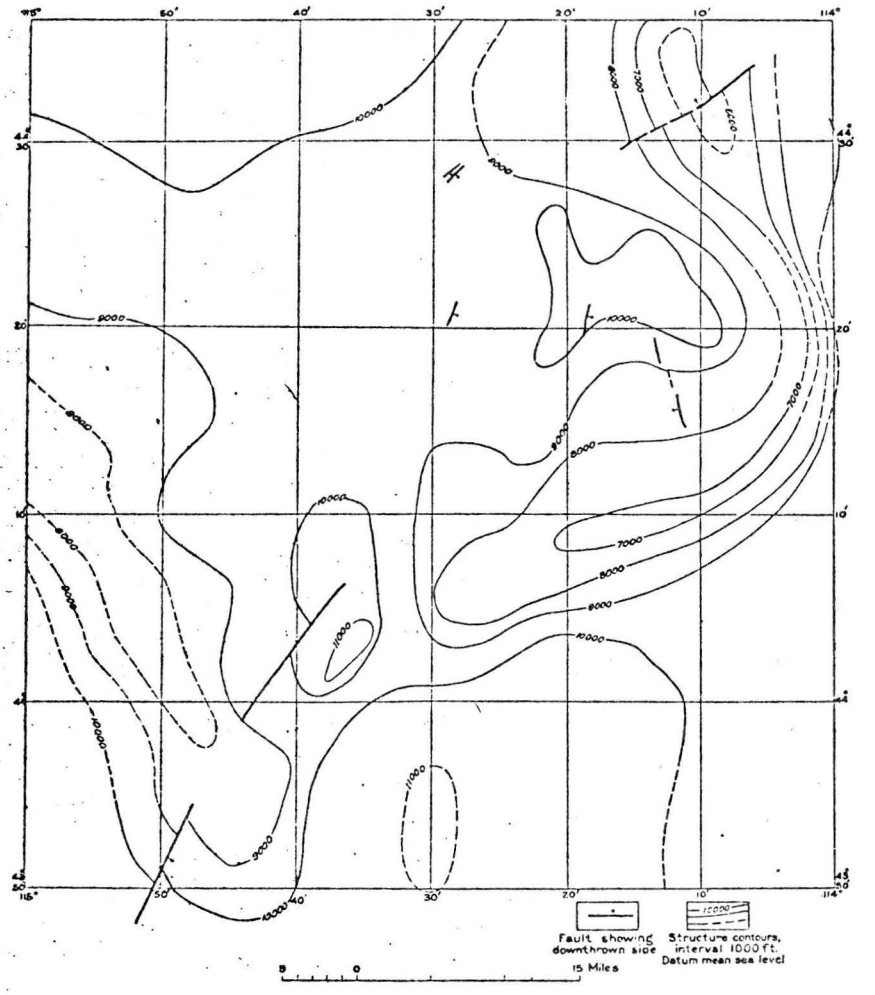


Figure 17. Hypothetical shape of the surface of the Challis volcanics after deformation but without erosion (Ross, 1937).

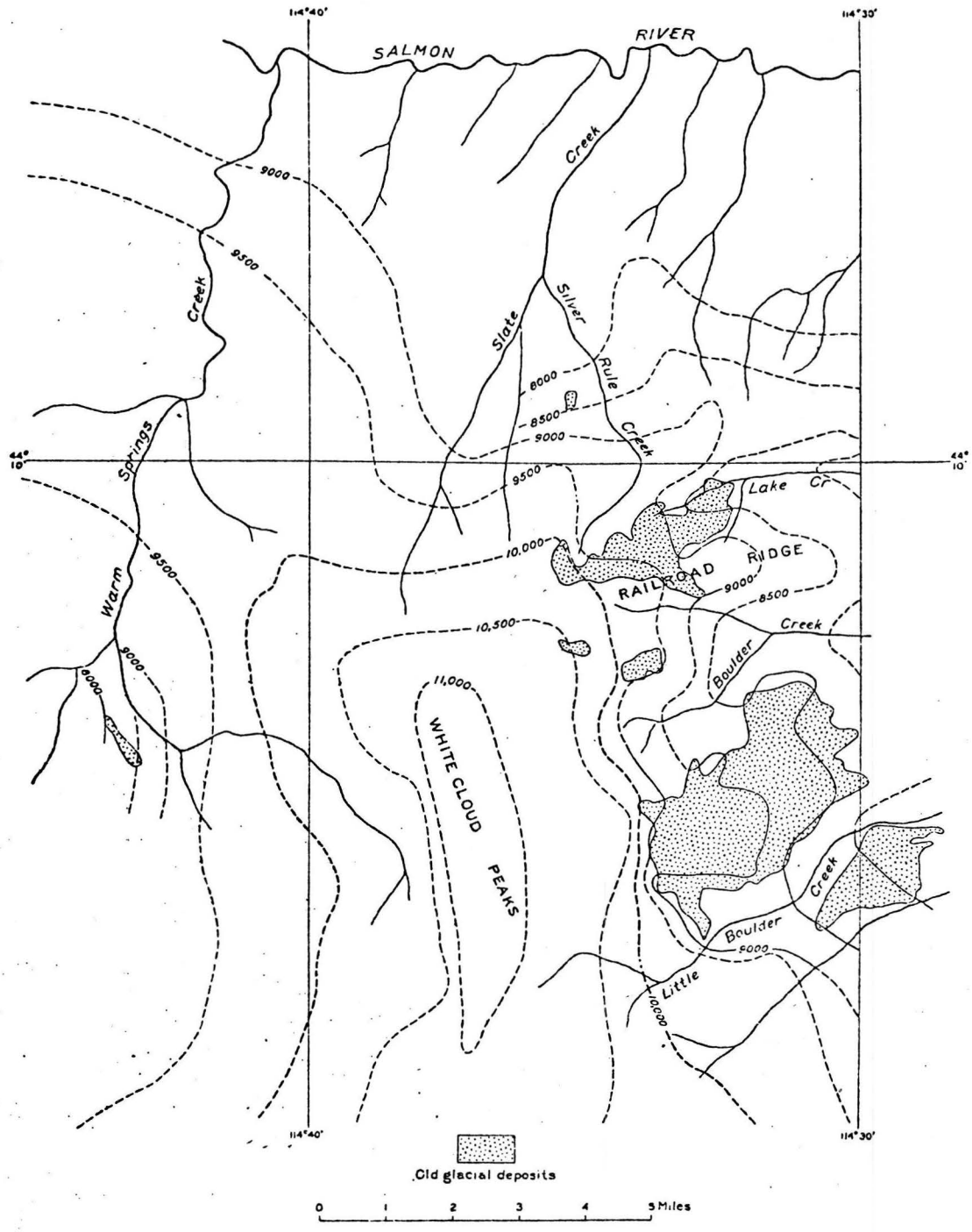


Figure 18. Restoration of the surface in the Bayhorse area on which the Early Pleistocene deposits were laid down. (Ross, 1937)

THE GEOLOGY OF SECTION III

SALMON UPLANDS SECTION

Section III covers almost all of the Elk City AMS map and the eastern quarter of the Grangeville map. The area around Elk City is characterized by rugged topography and includes the northern part of the Idaho Primitive Area, the Salmon River Breaks Primitive Area, and most of the Selway-Bitterroot Wilderness Area. The main drainage in the Section is the Salmon River system. Major lithologic units on the Elk City map include rocks of the Idaho batholith and a great complex of metamorphosed Precambrian Belt Supergroup rocks.

The portion of Section III on the Grangeville map is drained by the Salmon and Little Salmon Rivers. Major rock units in the area include the Idaho batholith, including border facies of the batholith, some Columbia River basalts, the Seven Devils volcanics, and various metamorphic rocks.

The Precambrian Belt Supergroup which is so important to the geology of the Elk City region and to all areas north of Elk City and beyond Idaho into Canada is discussed in the following section in some detail while the rock units in the Grangeville area are more conveniently discussed in the appropriate subsections.

The Precambrian Belt Supergroup

Covering most of the panhandle of Idaho, western portions of Montana, parts of eastern Washington, and extending into Canada, is a thick sequence of sedimentary-metasedimentary, Precambrian rocks called the Belt Supergroup. Harrison (1972) gave the following short discussion of the Belt terrane. Figure 19 and 20.

Belt Terrane

> "Sedimentary rocks of Belt age (about 850 to 1,450 m.y. old) are exposed locally from northeastern Alaska through Canada to southern Arizona and California (King, 1969a). In Canada, these rocks are called the Purcell Supergroup or System. The most extensive, continuous, and least disturbed exposures, however, are those of the northwestern United States and adjacent parts of Canada. Even though it is convenient to refer to this area as the 'Belt basin,' it is apparent that the basin is only an epicratonic re-entrant of a sea that extended along the western edge of the North American Precambrian craton during Belt time.
>
> Limits of the Belt basin are not clearly defined inasmuch as parts of it are covered by younger rocks, and other parts are virtually unknown. Belt rocks rest on Precambrian crystalline rocks east of Helena and are bounded by a fault that was active in Belt time (McMannis, 1963) in the area south of Helena. Geologic evidence (Price, 1964;

Mudge, 1970) and geophysical interpretations (Kleinkopf and others, 1968) indicated that the eastern edge of the Montana Disturbed Belt, which extends northwest from Helena, is along the approximate eastern edge of the Belt depositional basin. The Belt Supergroup is overlapped by the younger Precambrian Windermere System in the northwest, and it is buried under Columbia River Basalt in the west. The western and southwestern parts of the basin contain complexly deformed high-grade metamorphic Belt or pre-Belt rocks adjacent to, and as isolated patches within, the major batholiths. Correlation of these rocks with the Belt is largely by chemical similarity rather than stratigraphy.

The degree of metamorphism of Belt rocks increases across the basin from northeast to southwest and with depth in the stratigraphic section, as demonstrated by illite transformations to the 2M polymorph (Maxwell and Hower, 1967) and by an obvious increase in biotite content. Most rocks are greenschist facies, although those in the upper part of the section in the eastern part of the basin are almost unmetamorphosed, whereas those in the west and southwest may reach amphibolite facies, particularly near major plutons.

Belt rocks have been described in detail by many workers; only a general description of them as given by Harrison and Grimes (1970, p. 3) will be repeated here:

Belt rocks are generally monotonous in appearance because of fine grain size and drab color. The bulk of the supergroup has a grain size of silt or clay, and medium sand is the coarsest grain size observed in thousands of feet of rock over large areas. Quartzites and relatively pure to impure dolomites and minor amounts of limestones are scattered through the supergroup, although the carbonate rocks are uncommon in the lower part. Some carbonate rocks contain fossil algal forms, which are the only megascopic evidence of Precambrian life in the rocks. All rocks are dense and hard, reflecting the widespread low-grade metamorphism of most of the supergroup. Because of the metamorphism in most rocks,...we use terms such as argillite, siltite, and quartzite rather than claystone, shale, siltstone, and sandstone to describe the rocks.... Most rocks contain shallow-water structures, although the abundance of such structures decreases noticeably in the lower parts of the section. Disconformities in the supergroup are difficult to identify, and angular unconformities are rare.

Attention should also be called to data presented in that report (Harrison and Grimes, 1970) showing that Belt rocks from two widely separated areas are all graywacke or similar to graywacke and that the ratio among quartz, potassium feldspar, and plagioclase in siltite is almost

identical in rocks from both areas. New information confirms that the two stacks of rocks compared in that report had nonhomogeneous and different source areas, thus indicating a remarkable uniformity in weathering, transportation, and sedimentation processes affecting the two cratonic source areas during Belt time.

Facies changes in the Belt are very noticeable near the edge of the Belt basin in the Montana Disturbed Belt (Price 1964; M. R. Mudge, 1970, oral commun.) or even spectacular where the one major conglomerate in the supergroup was deposited (McMannis, 1963). But in most of Belt terrane the changes are subtle and require careful study across tens or even hundreds of kilometers to be defined. Even then, a facies change may involve only an increase in an abundance of argillitic siltite over silty argillite or in an abundance of medium- to coarse-grained siltite over very fine to fine-grained quartzite. The general monotony of the section and cyclical sedimenation, on scales ranging from millimeters to kilometers, reflect the cyclical recurrence of a few depositional environments and the consequent reappearance of remarkably similar rocks in many formations. Most attempts to study basin-wide sedimentation have understandably focused on the more unusual rock types, primarily the carbonate rocks (Smith and Barnes, 1966; McKelvey, 1968).

Belt terrane has been invaded by a variety of igneous rock types during Precambrian and Mesozoic-Cenozoic times. In general, the size and abundance of intrusive rocks of both groups increase from east to west and south. Principal Precambrian intrusive rocks are dioritic to gabbroic sheets known generally in the west as Moyie or Purcell sills, and simply called sills of late Precambrian age in the east. Small granitic plutons are known in southern British Columbia near the western edge of Belt terrane (Rice, 1941; Leech, 1962; Gabrielse and Reesor, 1964; Ryan and Blenkinsop, 1971). The only known Precambrian extrusive rocks are the Purcell Lava (Price, 1964) of southeastern British Columbia and adjacent parts of the United States. The Mesozoic and Cenozoic intrusive rocks range from trachyandesite sills in the eastern part of Belt terrane to granitic plutons in the south and west. Tertiary volcanic rocks cover extensive areas near the batholiths.

Major geologic and tectonic provinces involving Belt terrane are shown on Figure 20. Although it is not the purpose of this report to discuss at length the tectonic history affecting the basin, a few observations are pertinent. Most tectonic and geologic provinces and features reflect Cretaceous-Tertiary events, but a few are clearly remnants of Precambrian events. Precambrian movement along the Osburn strike-slip fault zone in the

western part of the Lewis and Clark line has been documented by Hobbs and others (1965); Precambrian movement along the Hope strike-slip fault seems reasonably clear (Harrison and others, 1972); and Precambrian movement along the basin-boundary Central Park (Willow Creek) normal fault has been documented by McMannis (1963). The Purcell anticlinorium probably reflects the late Precambrian East Kootenay orogeny of White (1959). Some faults within the Lewis and Clark line, along and branching from the Hope fault zone, and out in the block fault terrane, are also known to be Precambrian, as dated by lead isotopes from galena found along the faults (Zartman and Stacey, 1971). It also seems clear that the northeastern edge of the Montana Disturbed Belt reflects the approximate edge of the old Precambrian craton (Price, 1964; Mudge, 1970).

Superposed on the Precambrian features, and rejuvenating them at some places, are Cretaceous-Tertiary structural features that include three grabens--the Purcell trench, the Libby trough, and the Rocky Mountain trench--and a block fault terrane of basin and range type. Related to these in time and perhaps genesis are the Montana Disturbed Belt on the east and the Kootenay arc mobile belt on the west. The Cretaceous-Tertiary tectonic events will not be further discussed; because, as important as they appear to be, they have had little effect on the size, shape, and discernible sedimentation patterns of the Belt basin.

Thicknesses of the exposed Belt section in the basin as reported by various authors are shown in Figure 3. The thickest continuous section is near Alberton, Montana, where about 52,000 feet of Belt strata is exposed (J. D. Wells, 1971, oral commun.). An additional 15,000 feet of the lowest unit is exposed a few miles to the west (Wallace and Hosterman, 1956). Thus, the total exposed Belt section in one area, where neither the top nor base of the section is exposed, is 67,000 feet. The fortuituous culmination of the Purcell anticlinorium, which brings up the lower strata, combined with extensive gravens within the culimination, which drop down the higher strata, allows the thick section to be measured in low-dipping to horizontal strata in an area having a diameter of only about 50 miles." (Harrison, 1972).

The areas covered by the Belt and the total thicknesses at various locations are shown in Figure 21. Figures 22, 23, and 24 show the three cross sections represented by lines A, B, and C on Figure 21. Figure 25 gives the stratigraphic sequence and equivalent units for the Belt as suggested by Harrison (1972).

A double line at the southern edge of the Spokane and Wallace AMS maps marks very generally where metamorphism in the Belt becomes noticeable (zeolite facies or lower to the north of the line and

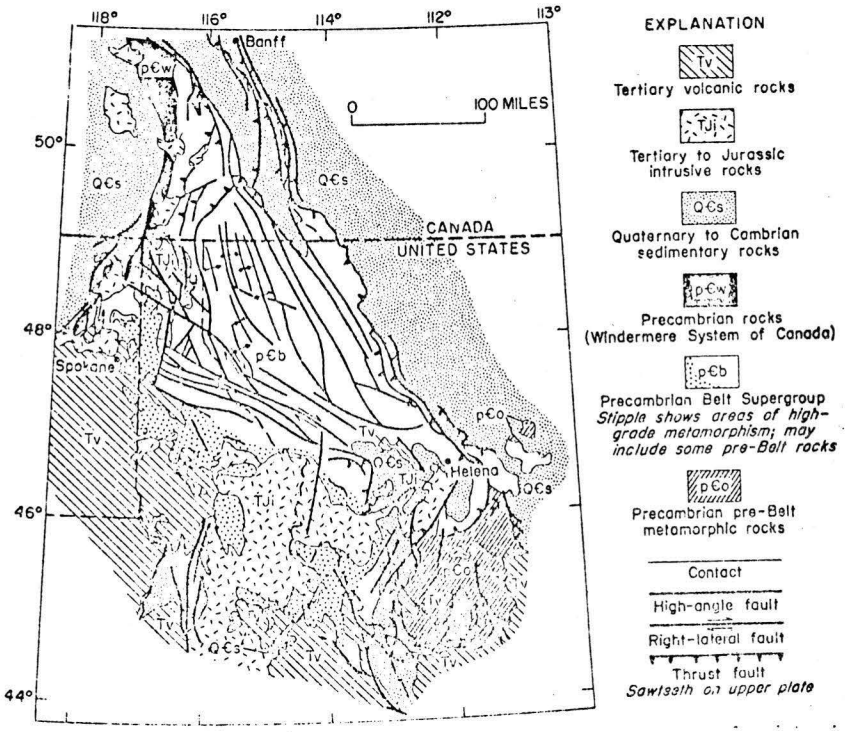


Figure 19. Generalized geologic map of the area of Belt terrane modified from Tectonic map of North America (King, 1969).

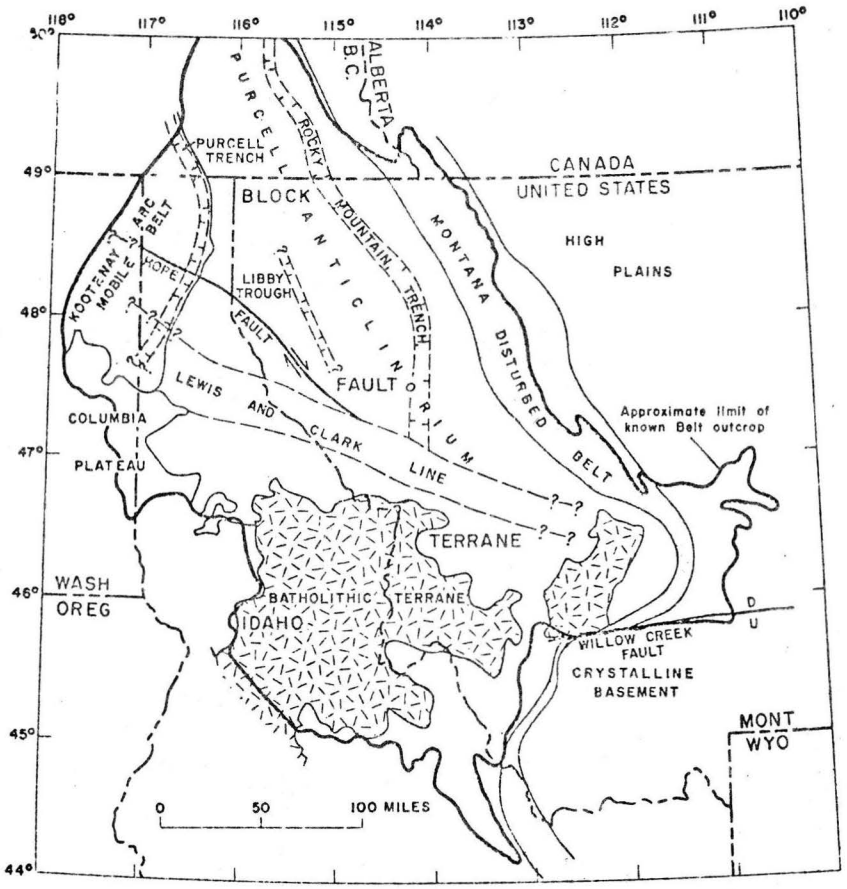


Figure 20. Principle geologic and tectonic provinces, eastern Washington, northern Idaho, western Montana, and Southern British Columbia U=upthrown side of fault, D=downthrown side. (Harrison, 1972)

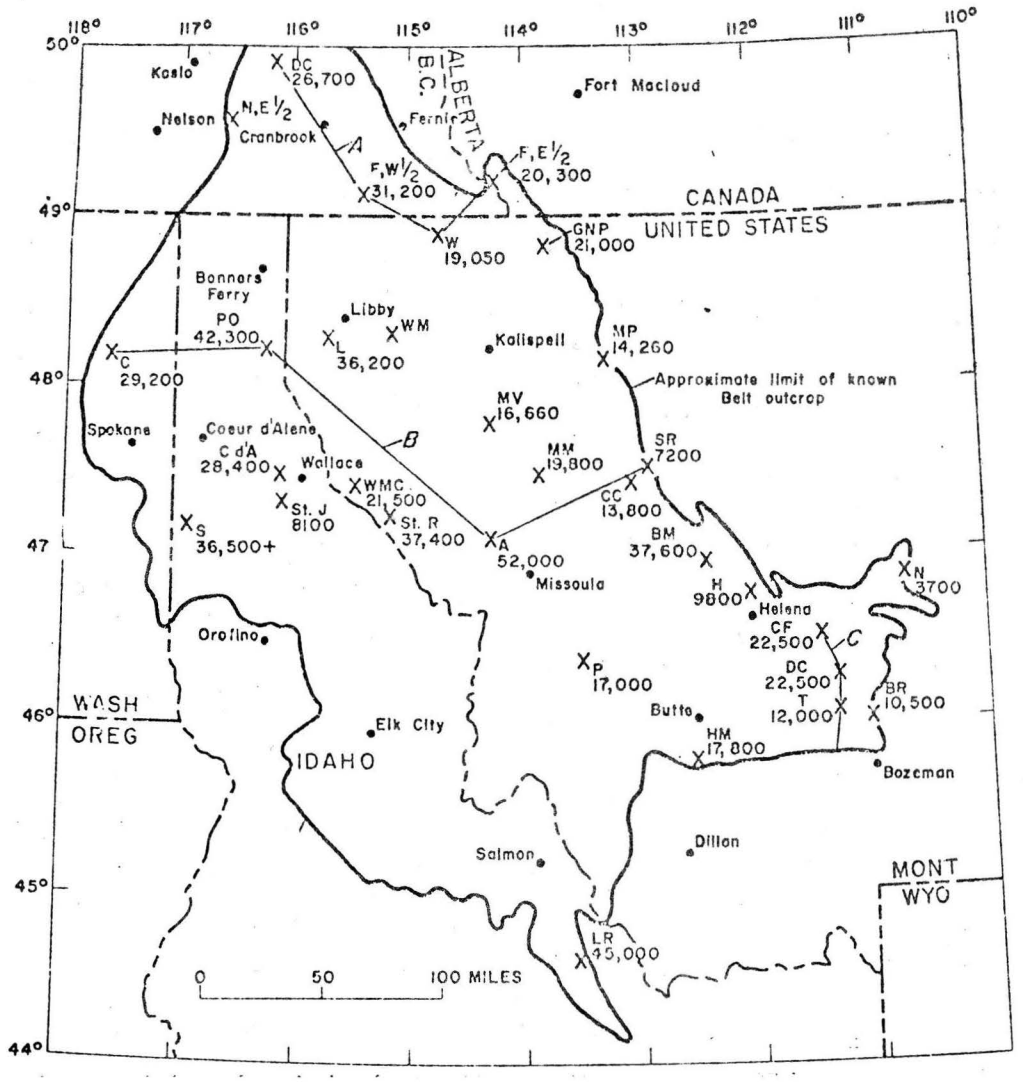


Figure 21. Thickness (feet) of Belt rocks exposed in various areas and lines of sections shown in Figures 22, 23, and 24. (Harrison, 1974)

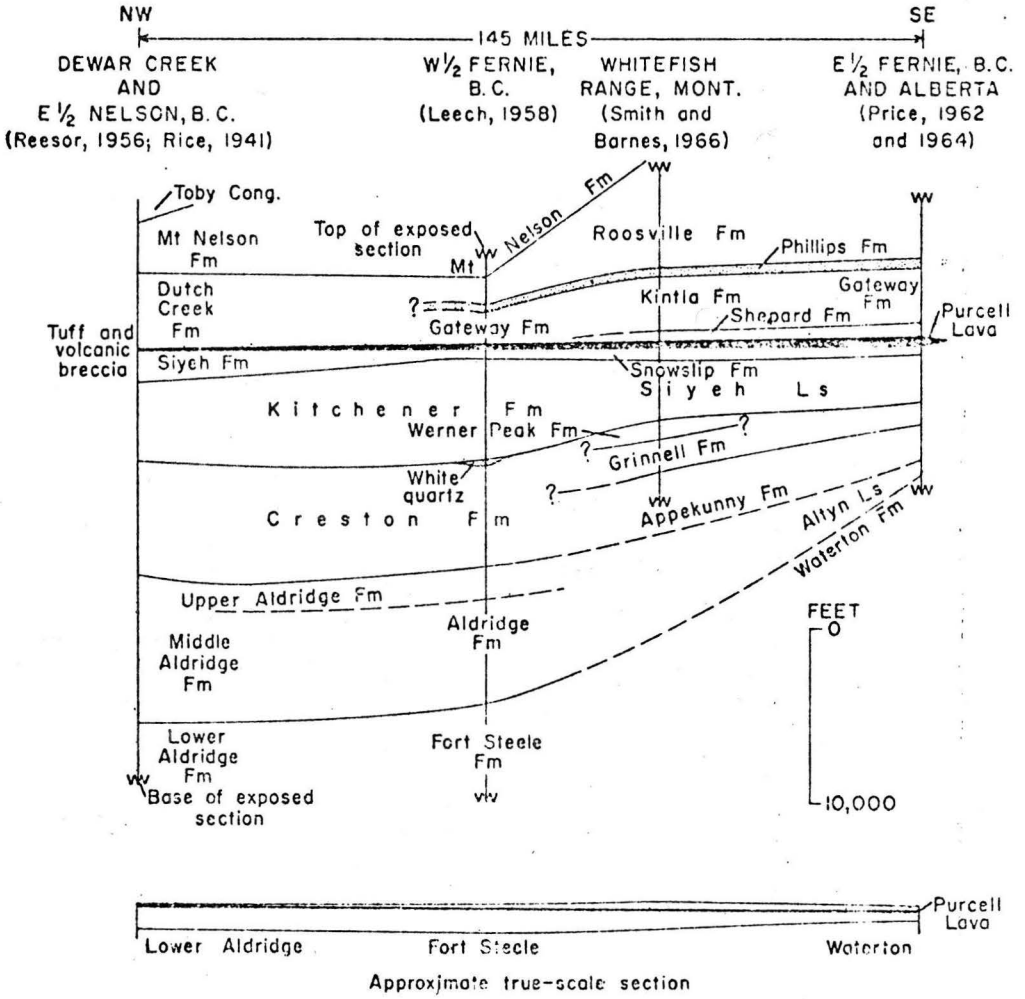


Figure 22. Belt cross section A (Price, 1964)

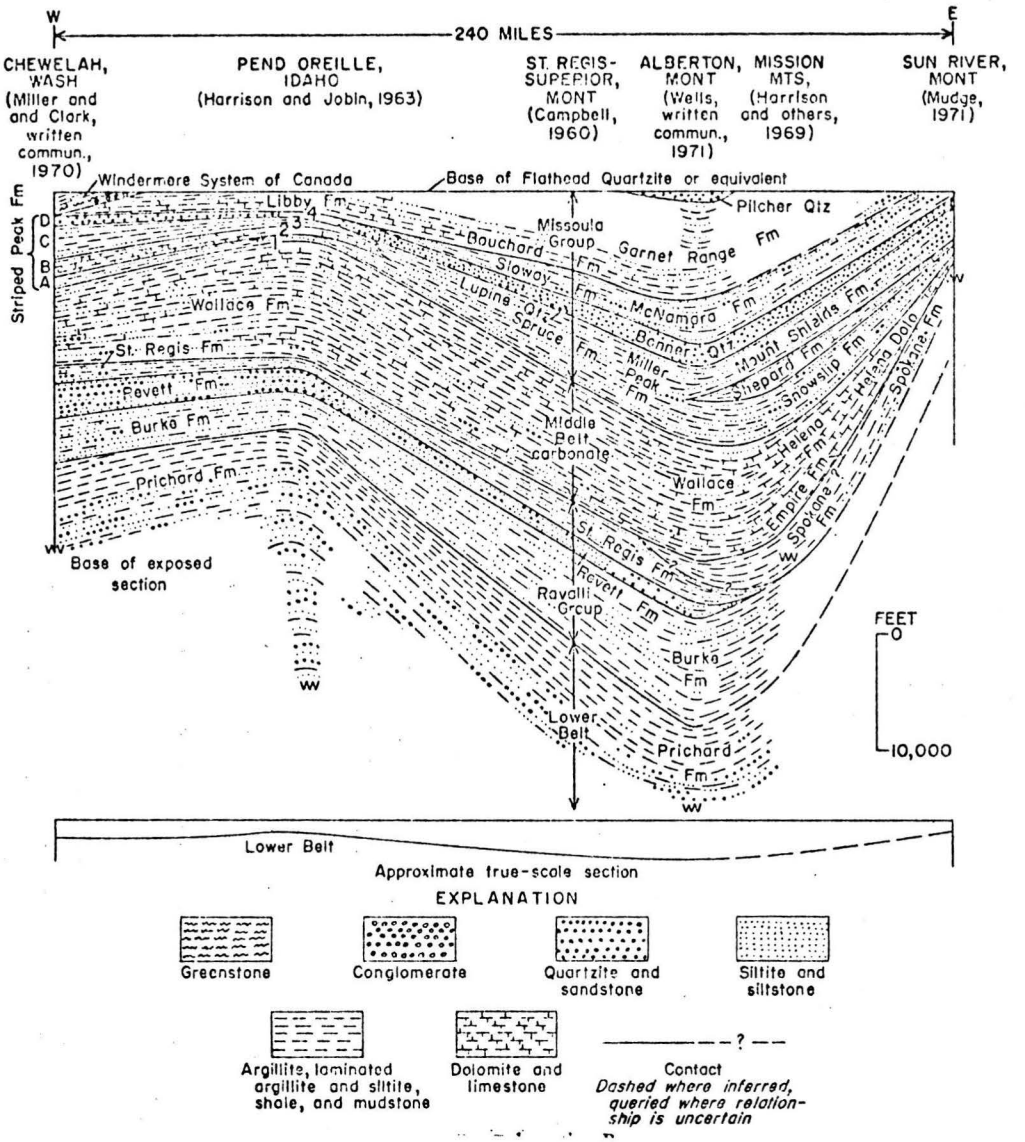


Figure 23. Belt cross section B (Harrison, 1972)

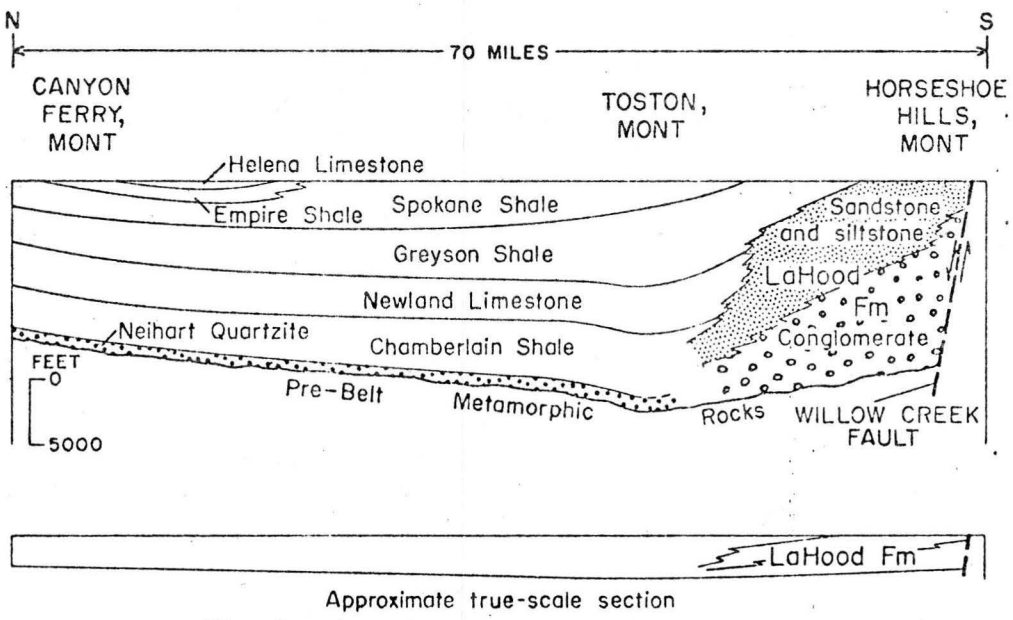


Figure 24. Belt cross section C (McMannis, 1963)

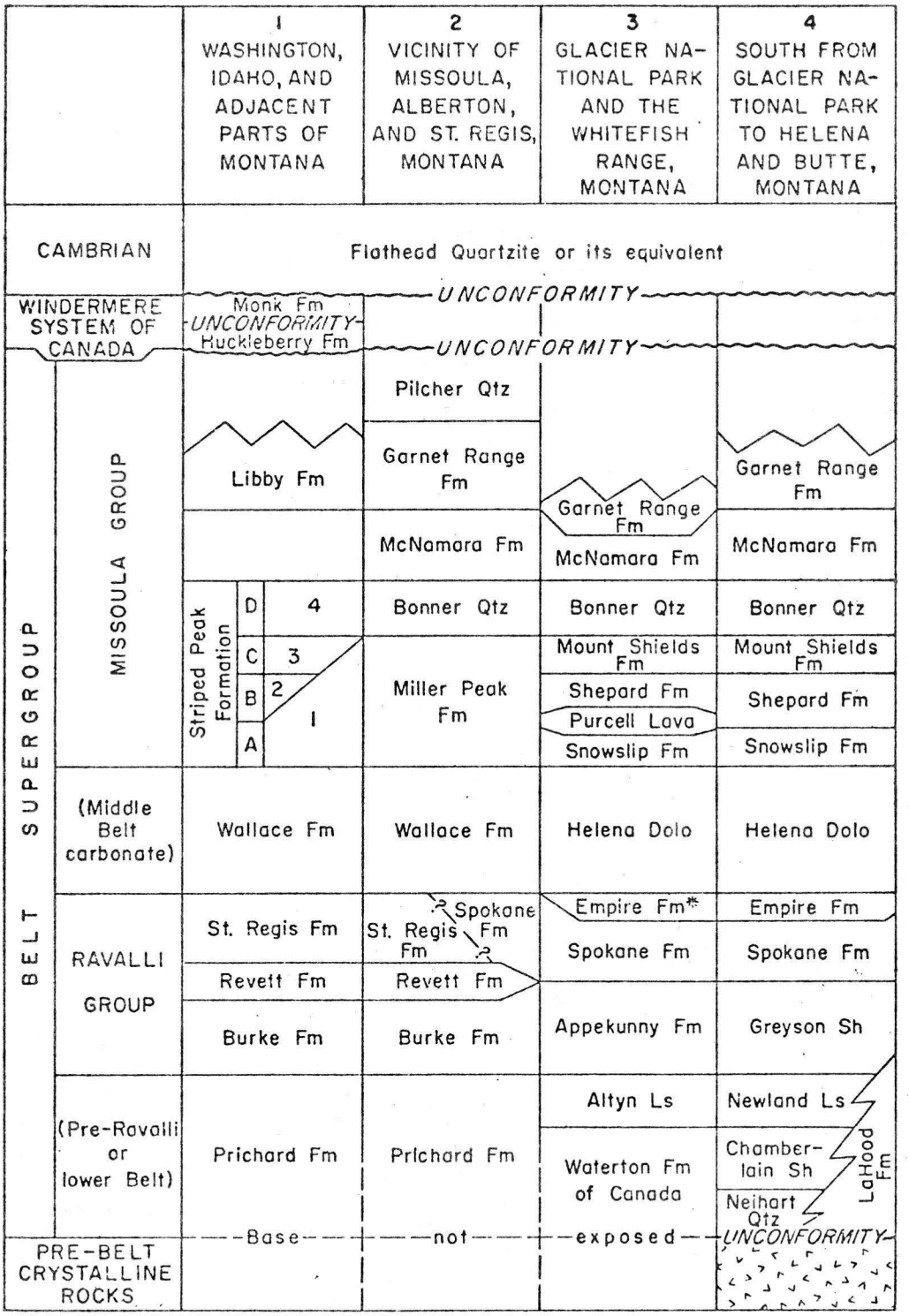


Figure 25. Recommended terminology for the Belt Supergroup. (Harrison, 1972)

greenschist facies or higher to the south of the line). There are numerous lithologic descriptions of Belt Supergroup units. The rocks of the Libby Formation through the Prichard Formation are described by Griggs (1972) as follows.

Libby Formation

"Dominantly medium-gray to olive colored siltite or laminated siltite and argillite; very thinly laminated dark argillite makes up the lowest part of section in northeast corner of map and parts of uppermost beds exposed in southwestern area of outcrop and also is interspersed at many other places. Several carbonate-bearing zones to several hundred feet thick occur in southwestern area; usually dolomitic siltite or argillite but includes some massive beds of blue-gray dolimite and some stromatolite layers. Contains rare chert laminae in the northeast, but none in the southwest. Mud chip breccia and ripple marks are common structural features; mud cracks are rare. Characteristically weathers in a blocky habit. Many of exposures in southwest altered to a punky, bleached rock of faded olive-green to rusty-tan color. Maximum thickness of eroded remnants in northeast is a possible 2,000 feet and in southwest maximum thickness varies from estimated 5,000 to 7,500 feet. Unknown amount of upper part eroded away. At some places in south is regionally metamorphosed to extent that secondary biotite is formed."

Striped Pack Formation

"Changes significantly from north to south both in thickness and lithology. At north a distinctive multi-unit consisting of a basal mixed siltite, argillite, and quartzite member of red and green color overlain successively by a tan dolomitic member, a very thinly laminated dark-gray argillite-siltite member and a dark red arkosic quartzite unit (Harrison and Jobin, 1963). Combined thickness totals near 2,000 feet and is about equally divided among the four units. Basal unit thickens and overlying units wedge out southward. South of Coeur d'Alene River predominantly gray to green to purplish red siltite with beds or layers of lighter colored quartzite; partings or interlaminations of darker argillite common; carbonate bearing rocks interspersed throughout section, but in minor amounts, usually as lenticular pinkish dolomite segregations, but occasionally as dolomitic bed or relatively thin zone. Medium- to coarse-grained, light-gray to reddish, feldspathic quartzite 500 feet plus thick forms a persistent zone throughout the southwestern area of exposure; it occurs about 1,000 feet below the top of the formation. Mud cracks, ripple marks and mud chip breccia in red to green colored rocks; salt casts

and channeling are much less common structures. Several stromatolite layers occur in northern exposure areas. Micaceous sheen on bedding surfaces is characteristic at most outcrops. Altered to a punky and bleached rock of faded red to cream or rusty tan color at many outcrops in southwestern area. Maximum thickness at southern edge of map 5,000 feet plus. No thick transition zone into overlying Libby Formation, or into underlying upper part of Wallace Formation is evident."

Wallace Formation

"Undifferentiated in Medical Lake area southwest of Spokane, where occurrences are poorly exposed in isolated hills that stand up above the top of basalt surface, and usually show a wide range of effects from metamorphism. Varies from gray, fissile, laminated argillite or phyllite to calc-silicate hornfels or granofels, sugary quartzite, mica schist, biotite-quartz plagioclase gneiss and migmatite."

Upper Part of Wallace Formation

"Very thinly-bedded black argillite or thinly laminated black argillite and light gray siltite; occasional bed or thin zone of lighter colored siltite scattered through unit. Carbonate-bearing zone several hundred feet thick near center or towards the top and appears persistent throughout area; consists of greenish-gray to gray to dark-gray interbedded to interlaminated dolomitic argillite to siltite with some gray dolomite to dolomitic limestone beds; similar to basal unit of the lower part of Wallace. At some places black argillite, also is carbonate bearing. At most exposures rocks are noticeably fissile, and fairly regularly bedded. Along southeastern margin of map metamorphosed, and changed to a phyllite to mica-garnet schist; some of which contain abundant scapolite. Thickness 2,500 to 3,000 feet. Lower contact graditional."

Lower Part of the Wallace Formation

"Two distinguishable members, here undifferentiated. Lower unit, green to greenish-gray or gray, usually carbonate bearing, interbedded or interlaminated argillite and siltite; contains many bluish-gray dolomite and dolomitic limestone beds. Blocky weathering, molar tooth structure, and rusty tan weathering rind are characteristic. Upper unit consist predominantly of alternating black argillite and light to greenish-gray siltite or quartzite, latter usually carbonate bearing; near southwestern margin quartzite and siltite becomes dominant in zones hundreds of feet thick. Interspersed in sequence are layers or zones of rock like that in the

Wallace units above and below. Mud cracks (non-desiccate in origin) and ripple marks common; fine textured cross bedding evident on etched surfaces of some siltite and quartzite beds. Irregularity in bedding and minor folds characteristic. Along southeast margin of map metamorphosed to a hornblende--or diopside--plagioclase hornfels to granofels or gneiss and mica schist; in part scapolite bearing. Thickness estimated to vary from 5,000 to 7,000 feet; thinnest in and around Coeur d'Alene District."

St. Regis Formation

"Dark red, purplish red, to green or greenish-gray; interbedded or interlaminated usually very thin- to thin-bedded argillite and siltite; contains some quartzite beds in basal part and becomes more argillitic toward top. Some carbonate-bearing beds in upper part. Mud cracks, mud chip breccia, and ripple marks very common. Thickens from less than 1,000 feet at north edge to about 2,000 feet in southern part of map. Graditional into units above and below. Weathers into platy or flaggy fragments."

Revett Formation

"Blocky, white to light gray, thick-bedded, fine- to medium-grained, vitreous quartzite; somewhat feldspathic. Gray to greenish-gray, thin- to thick-bedded siltite with partings and interbeds of argillite common in upper and lower parts; siltite may be dominant rock types over zones tens of feet thick. Cross bedding in vitreous quartzite common and a rusty speckling due to weathering of small, round carbonate segregations in some vitreous beds characteristic. Thickness from 2,000 to 3,000 feet. Transitional over hundreds of feet into units above and below."

Revett and Burke Formations, Undifferentiated

"Units lumped together where identity questionable in metamorphic terrain south of St. Joe fault at east edge of map, or in areas west of south end of Coeur d'Alene Lake and south and west of Spokane where rocks are poorly exposed in isolated hills and ridges. Metamorphosed rocks vary from white, coarse-grained, vitreous quartzite, to micaceous quartzite and mica schist. In isolated hills mostly light to greenish-gray siltite to white vitreous quartzite, a few exposures are of rock metamorphosed to micaceous quartzite or mica schist. Some outcrops bleached and punky."

Burke Formation

"Light- to medium-gray to greenish-gray, thin- to thick-bedded siltite with partings and interbeds of argillite. Some light-gray to white quartzite in scattered beds

and zones. At some places, mostly peripheral to Coeur d'Alene mining district, and in the lower middle part of section rocks are reddish-purple to lavendar in color. Ripple marks are common in places; some cross bedding. At many exposures has a faded weathered rind that contrasts with darker fresh rock. Fine magnetite octahedra pepper many of siltite beds. Transitional into formations above and below."

Prichard Formation

"Undifferentiated where the formation was mapped as a single unit in western part of Coeur d'Alene district near Kellogg, and also where rocks are highly metamorphosed as those northeast to southeast of Spokane and in the southeast corner of the map area. The upper transitional zone and argillitic rocks have not been mapped separately from the predominantly siltite to quartzite below in the western part of the Coeur d'Alene district. The metamorphosed rocks consists mainly of mica schist, quartz- plagioclase-biotite geniss and quartzite. In area east of Spokane gneiss, medium- to coarse-grained and in greater abundance than schist; here also sillimanite widespread and a decrease in metamorphic grade below sillimanite isograd noted only at northern margin and in peripheral rocks to south and southwest; migmatitic rocks also quite common in this region and at places underlie large areas such as along southern margin of Spokane Valley near Post Falls, and on south from the vicinity of Dishman. In this area includes the Hauser Lake Gneiss (Weis, 1968). Mica (mostly biotite) schist is more common than gneiss in metamorphic rocks in southeast corner; kyanite is a minor but common constituent in many layers and small amounts of garnet are almost ubiquitous; migmatite, in irregular zones to veinlets, a minor constituent (Heitanen, 1963b)."

Upper Part of Prichard Formation

"Black to gray, very thin-bedded argillite commonly interlaminated with light gray siltite and also containing some siltite beds. Grades upward into interbedded and interzoned argillite, siltite, and quartzite sequence, which forms the transition into siltitic and quartzitic units above and which is from about 500 to 1,000 feet plus thick. Argillite rust stained on weathered surfaces; occasional ripple marks in upper part. Total thickness from 2,500 to 3,500 feet."

Lower Part of Prichard Formation

"Predominantly medium- dark-gray, thin- and regularly-bedded siltite, laminated in part, some argillite in laminae and beds. Some beds or zones of gray to white quartzite of lenticular habit. Disseminated pyrrhotite

concentrated within certain laminae characteristic, and its weathering results in a persistent rusty red rind on fracture surfaces. Thickness 7,500± feet; base not exposed. Medium- to coarse-grained, in part vitreous, but mostly micaceous quartzite; in zones to hundreds of feet thick with some interlayered more micaceous rock. Shown separately only in the more highly metamorphosed rocks. A variety of gneiss in which feldspar porphyroblasts up to two inches in length are characteristic; cataclastic features common throughout; shears and mylonite zones roughly parallel foliation of gneiss. It has been named Newman Lake Gneiss (Weis, 1963)."

Section III contains only metamorphosed Belt rocks. The correlation of various metamorphic units with Belt equivalents is tenative in the central and southern portions of the Elk City map because of increasing metamorphic grade. Only three members of the Belt Supergroup are recognized in the metamorphic terrane, the Prichard Formation, the Ravalli Group and the Wallace Formation. Reid, Morrison, and Greenwood (1973), have summarized the complex history of metamorphism in the Belt as follows. (Table 2)

General Considerations

"Rocks of the Belt Supergroup in central and north Idaho were involved in the orogenic events that produced the extensive Cordilleran foldbelt, including intrusion by Mesozoic and Tertiary batholiths. Near Mesozoic to Tertiary intrusive centers the Belt rocks are metamorphosed to high-grade gneiss and schist and show structures indicating a complex history of plastic flowage and cataclasis. Away from the intrusive centers the Belt rocks appear to be typical supracrustal rock which in Montana unconformably rest on a crystalline pre-Belt basement metamorphosed in Hudsonian to Kenoran age orogenies (King, 1969).

Phanerozoic units overlap the Belt Supergroup and rest unconformably on pre-Belt basement in Montana. The structure and stratigraphy of the part of the Belt Supergroup away from intrusive centers is of economic interest because of the mining in these areas and has been reported in considerable detail (summarized by Ross, 1963; Harrison and Campbell, 1963, 1968; Smith and Barnes, 1966; and Harrison, 1972). The lack of an economic incentive provided by active mining and rugged terrain has limited detailed mapping in the highly metamorphosed parts of the Belt near the Idaho batholith. So little was known of this region that Harrison and Campbell (1963) indicated that the original southern boundary of the Belt depositional basin was probably at or near the present northern contact of the Idaho batholith as it was then shown on the state

geologic map (Ross and Forrester, 1947). A recent review (Harrison, 1972) extends the southern boundary of the Belt basin much farther south. Geologists who have mapped, compiled, or reviewed the geology of the border rocks of the Idaho batholith have generally been struck with the proximity of the highly metamorphosed Belt to the batholith rocks. A consensus developed that orogenic folding and metamorphism of the Belt Supergroup in this area first occurred in Mesozoic time (Ross and Forrester, 1958, p. 32-33; and King, 1969, p.44). Harrison (1972, p. 1217-1218) states that most of the tectonic and geologic provinces and features involving Belt rocks result from Cretaceous to Tertiary orogenic events. He lists Precambrian movement on the Osborn, Hope, and Wilson Creek faults and late Precambrian folding of the Purcell anticlinorium as exceptions to this rule. Hietanen (1968) reposts that the Belt Supergroup north of the Idaho batholith underwent two periods of metamorphic recrystallization in the Mesozoic. The first period was synkinematic to the major folding of the area. The second period was post-kinematic and slightly hotter than the first. She relates these periods to a two stage Mesozoic development in the Idaho batholith. During the synkinematic stage, magma was formed at depth. The emplacement of this magma at the presently exposed level in a post-kinematic stage caused elevated temperatures and produced a static recrystallization.

However, the reported spatial relation between the Mesozoic batholith intrusion and tectonism and metamorphism of the Belt Supergroup recently has become much less convincing. A more or less continuous intrusive zone of Mesozoic batholiths has been reported connecting the Idaho batholith to the Sierra Nevada batholith to the southwest (Taubeneck, 1971) and to the Boulder batholith to the north (Hyndman and others, 1972). The great heterogeneity of the intruded rocks in this zone, as to age, structural history, and degree of metamorphism, makes a simple relationship of a single synkinematic metamorphic recrystallization followed by upward emplacement of magma generated during the metamorphism highly unlikely.

It should be observed that the Mesozoic batholith zone is highly discordant to the dominant northwesterly structural trend in the Belt rocks (north of the Idaho batholith) reported by Hietanen (1968) and Reid and Greenwood (1968). Strong discordance of batholith contacts with metamorphic structure is also observed on maps of Hietanen (1962, 1963a, 1963b, 1963c, and 1968) and in outcrop. Moreover, the simple picture of one major deformation is not sustained by our field observations. Complexities of the structural history of the area have been presented (Reid and Greenwood, 1968) and will only be summarized here. Evidence of multiple fold events is to be found throughout the Belt

terrain north of the Idaho batholith: Schistosity is isoclinally folded; fold axes and axial surfaces are folded, garnets contain platy quartz inclusions that indicate multiple episodes of growth and rotation, orthogneisses which intruded and were cleaved late in one fold event were isoclinally folded by a succeeding fold event; migmatitic layering is isoclinally folded and cross-cut by granitic stringers parallel to the axial surfaces of the new folds. Two generations of folds can be observed in most outcrops and it is not uncommon to find three or four sets of crossing lineations in a single outcrop.

If a simple structural picture is not sustained, what then about the reported synkinematic and post-kinematic recrystallization events? Here again, the picture is more complicated than it appears. Reid and Greenwood (1968) report three periods of metamorphism for the Belt rocks, each of which had synkinematic and post-kinematic phases. These periods are separated by structural events, as well as by magmatic events in the higher grade zones. The first recrystallization period produced maximum assemblages including sillimanite plus muscovite and sillimanite plus kyanite. Magmatic intrusions during this period included: gabbro, now metamorphosed to amphibolite; granodiorite, now metamorphosed to augen gneiss, and extensive formation of migmatite by processes of segregation and injection. The second period of recrystallization produced maximum assemblages of sillimanite plus orthoclase (without muscovite) and sillimanite plus andalusite. Magmatic intrusion during this period included: large plutons of quartz diorite to granodiorite, now metamorphosed to schistose and banded gneiss; sills of diabase, now metamorphosed to amphibolite, and a second generation of magmatite which cuts folded layers of first generation magmatite. The third period of recrystallization produced maximum assemblages in the greenschist facies. Magmatic intrusion during this period emplaced plutons of quartz diorite to granodiorite, now moderately to strongly cleaved but not extensively recrystallized.

All of the above listed rocks are discordantly intruded by the several intrusive phases of the Idaho batholith. Although Cretaceous-age primary gneiss, near contact dike emplacement, and moderate cataclasis make the western contact somewhat diffuse on the map scale, the separate rock types have distinct contacts which can be recognized in the outcrop. The eastern contact of the batholith appears to lack primary gneiss.

The observed cycles of tectonism, metamorphic recrystallization, and magmatism are difficult to reconcile with the concept of a single orogeny in Mesozoic time. The

cooling and unroofing which produced the different stability fields of the recrystallization periods must have required considerable time.

To summarize the geologic evidence, three cycles of tectonism, metamorphic recrystallization, and magmatism preceded the intrusion of the Idaho batholith. These three orogenic cycles operated under such different conditions of temperature and pressure that considerable time seems to be required to allow for unroofing and cooling between the thermal pulses. The apparent time separation between the events suggests that they were produced by three separate orogenies.

As an independent test of the time separation of the proposed orogenic events and to determine their relation to reported orogenic ages in North America, samples of orthogneiss, high grade paragneiss, amphibolite, and migmatite have been subjected to isotopic age analysis using Pb-U, Rb-Sr, and K-Ar techniques.

Table 2 summarizes our concepts of the history of the Belt Supergroup in the area north of the Idaho batholith, using the format and data from Harrison (1972). Regional age orogenic names are from King (1969) and Roberts (1972).

The high-grade metamorphic Belt rocks appear to be part of a relict Precambrian fold belt reworked in the formation of the Cordilleran fold belt. The term Clearwater orogenic zone (Reid and Greenwood, 1968) has been proposed for the complexly metamorphosed Belt rocks. We now propose that the term should apply only to moderate to high-grade relicts of the Precambrian metamorphic fold belt. Low grade parts of the Belt, such as in the Coeur d'Alene mining district, were also metamorphosed in this period; however, the structural features of the low grade Belt appear to be dominated by Phanerozoic fold trends. The term Elk City event (Harrison, 1972) could be used to refer to the orogenic events which span the 1500-1150 m.y. period." (Reid, Greenwood and Morrison)

Subsection 1 - Granite Mountains Uplands Subsection

Subsection 1 is on the Grangeville AMS map. The subsection is bordered on the north by the Salmon River and on the west by the Little Salmon River. This area is included in the Riggins 30 minute quadrangle mapped by Hamilton (1969), Subsection 1 is in the border zone of the Idaho batholith. Gneissic quartz diorite (mKi2a) and gneissic quartz monzonite (mKi3a) make up most of the rocks in the area with Columbia river basalt (TeiC) and some glacial deposits (Qg) as minor units.

Table 2. Depositional and orogenic history of the Belt Supergroup in central and north Idaho after Reid, Morrison, and Greenwood (1973).

Geologic age (m.y.)	Regional unconformities	Sedimentary Deposits & Thickness	Magmatism	Tectonism	Metamorphism	Regional Orogenic age
<40				Regional cataclasis and uplift on Bitterroot Front		Basin-Range
40			Red granite of upper Selway River			Laramide
100			Idaho batholith			Coast Range
170			Cleaved granodiorite	Northeast-trending folds, left lateral wrench faults and related thrusts	Low temperature metamorphism	Nevadan
300				Cataclasis	Hydrothermal alteration	Antler
600	Paleozoic					
700		Windemere system of Canada (22,000+ ft.)	Volcanics gabbroic sills			
850	Windemere		Gabbroic sills Clearwater Ridge orthogneiss	East Kootenay orogeny, upright isoclinal folding on north-west axes	Low pressure - high temperature metamorphism, migmatization	Grenville
1100	Basal Missoula Group	Missoula Group (30,500+ ft.)	Purcell lava gabbroic sills	Local unconformities from basin instability		
1200			Couer d'Alene lead and uranium veins(?)	Folding and faulting in Couer d'Alene area	Sericite formation in Couer d'Alene area	Elsonian
1300			Granodiorite at Hellroaring Creek		Regional metamorphism in Prichard area Alberton, Montana	
1500			Augen gneiss of Red River and Priest River	Recumbent isoclinal folding	High pressure - high temperature metamorphism	
1700	Prichard	Middle and lower Belt (55,000+ ft.)				
>1700				Pre-Belt of Montana		

The quartz diorite gneisses grade from strongly foliated in the western portion of 1 to less and less foliated eastwards. This is typical of the batholith border zone. A core of gneissic quartz monzonite extends north-south through the central part of the subsection and a series of north-south faults have progressively downdropped blocks of rock going to the east. The obvious north-south trend of structure and rock units is common on the western margin of Idaho and has been mentioned several times before. Subsection 1 is quite different in lithology from any other subsection on the Grangeville map, which includes volcanics and metamorphic rocks.

Subsection 2 - Burgdorf Basin Subsection

Subsection 2 contains the Warrior Mining District and although an area of steep relief, includes two north-south trending valleys, the Secesh and Warren basins. The major drainage in the area is the Secesh River and its tributaries which flow into the South Fork of the Salmon.

Lithologies consist of small patches of quartzites, gneisses and schists which were believed by Reed (1937) to be Belt equivalents, quartz monzonite of the batholith which covers most of the area, and a few Tertiary dikes. Reed noticed two periods of glacial activity based on terrace gravels and valley fill material. Warren valley was described as an extensive alluvial plain which was covered with clay, sand derived from batholith rocks, and peat beds.

Reed (1937), described several joint sets in the subsection; a major set which trended N 80° E and a minor set trending N 10° W. He also mentioned that the few faults in the area were alligned north-south.

The formation of the various north-south trending valleys in this area is related to the "Old Erosion Surface" which is mentioned frequently in the literature. Reed summarized his ideas about the effects of the erosion surface on this part of Idaho as follows.

Tertiary Structure and Development of Topographic Features

> "Many of the Tertiary and younger structural movements in north-central Idaho, including the Warren district, were of such magnitude that they are responsible for some of the broader features of the topography. Therefore, Tertiary structural movements and the development of the topography are discussed under one heading.
>
> One of the most outstanding topographic features of central and north-central Idaho is the extensive but dissected surface or plateau that constitutes the locally termed "high country" in this region. This is an erosion surface and represents an old topography developed at a much lower altitude than that at which it

stands today. It is the so-called "summit" or Idaho peneplain and its extent, age, and manner of development are controversial.

Strictly speaking, this surface is not a peneplain at all, but rather a mature erosion surface of considerable relief, in many places as much as 500 feet within a mile.

Although little is known of the drainage system over this old surface except that most of the streams probably flowed in broad, locally gravel-filled valleys, some evidence contributed by the streams now flowing over remnants of the surface and fragments of old, gravel-filled channels indicate that the major drainage lines may have flowed either north or south roughly parallel to the trend of the rocks. Lake Creek and Secesh River, for example, flow southward away from the Salmon River Canyon for many miles. Old, gravel-filled channels high on War Eagle Mountain, and, according to local reports, others farther west along some of the eastern tributaries of the Little Salmon River, also indicate a southward trend of the drainage on the old surface.

Southwest of the Warren district near McCall and west and northwest of the western part of the Nezperce National Forest north of the Salmon River the Columbia River lava flows rest on the old erosion surface, thus indicating that it is older than Lower Middle Miocene. Toward the southeast, in the vicinity of Edwardsburg and Thunder Mountain, and beyond, the surface truncates rocks of the Challis volcanics, which have been determined as Lower Miocene. Thus, the erosion surface must have developed in the latter part of the Lower Miocene or very early in the Lower Middle Miocene.

The large areas of the surface developed on the quartz monzonite of the Idaho batholith, which must have crystallized at depths of several thousand feet, indicate that a large amount of erosion was involved in the cutting of the surface.

Following the formation of the erosion surface, a period of uplift resulted in raising it to its present altitude. One of the accompanying results of the uplift was the tilt of the surface toward the northwest. The tilting explains the generally higher altitudes of the mountain tops formed by the surface in the southeastern part of the Salmon River Mountains than in the western part of the Clearwater Mountains.

The elevation and tilting resulted in the development of major drainage systems (parts of the Salmon and Clearwater River systems) across the mountains in a westward

direction in disregard of the rock structure. Because of the relatively great elevation of the surface these westward-flowing streams cut deeply and rapidly, and their valleys progressing headward (eastward) captured the southward, and perhaps locally northward, flowing streams of the old drainage system across the erosion surface. An example of such stream piracy is the capture of the Secesh River by the South Fork of the Salmon.

As might be expected, the uplift and tilt of such an extensive block of the earth's crust as that embraced by the Salmon and Clearwater mountains were not performed without many local structural movements. Block faulting along the western side of both mountain groups and the western part of the Sawtooth Mountains around Boise Basin is common.

Small areas, many of them less than 20 miles long in a northward direction and about half as wide, within the mountains were elevated less than the surrounding mountains (were relatively depressed) and now form structural basins from which has come most of central Idaho's placer gold. Many of the basins are bounded on one or more sides by normal faults, most of which trend northward. Where not bounded by faults, the old erosion surface appears to be carried down into the basins by warping.

Most, if not all, of the structural movements attending the elevation and tilting of the old erosion surface and the formation of the basins were completed before the Pleistocene. This is shown by the widespread distribution of undisturbed moraines in some of the structural basins and in canyons cut below the old erosion surface. The moraines, such as that along Secesh River near where it is joined by Lake Creek, in the basins belong to an older glacial stage. The younger moraines--those found in the canyons such as the one in the upper part of the valley of Slate Creek in the Nezperce National Forest, represent the Wisconsin stage of the Pleistocene epoch.

The Warren Basin is representative of the structural basins described above. Along its western side outside the mapped area it is bounded by a northward-trending normal fault, on which the east side is the downthrown side, that forms the scarp bounding on the west Republican Flats and the flat divides between Houston and Shissler Creeks, and between Shissler and Thomas Creeks. The scarp is a conspicuous feature from the brink of the Salmon River Canyon near Cottontail Lookout southward to the valley of Thomas Creek. From there its projection would extend up the valley of Stratton Creek, but

it apparently has died out before it reaches the higher altitude of the vicinity of Steamboat Mountain.

No faults are shown on the accompanying geologic and topographic map of the Warren district, although the presence of at least three of the type just described is postulated from physiographic evidence. The bedrock scarp rising steeply from the western edge of the broad but low gravel terrace just west of Warren Meadows probably marks a fault. It trends west of north, and possibly joins the longer fault of greater displacement described above somewhere in the valley of Thomas Creek, where the topographic expression of the faults is absent. This fault was not recognized south of Steamboat Creek.

A smaller but otherwise similar fault, which probably has a displacement of less than 25 feet, may separate the low terrace mentioned in the preceding paragraph from the Warren Meadows. Test-pitting and drilling both on the Meadows and on the terrace, dredge operations, and placering at McGovern's and Hackney's pit all show that the bedrock floor beneath the gravels of the Meadows rises abruptly to the bedrock floor beneath the gravels of the terrace. The tilted sediments at McGovern's and Hackney's pit may reflect movement along this fault.

The surface formed by the local divides east of Warren Meadows projected westward across the Meadows would pass probably between 50 and 150 feet above the floor of the basin. A northward-trending fault, downthrown on the west, along the east side of the Meadows may account for this.

From the high surrounding divide on all sides of the basin the old erosion surface dips conspicuously inward except where the depression has been by faulting, as already described, and not by warping.

Evidence of relative movement between blocks bounded by joint planes is furnished by the striations observable on many joint surfaces. A large amount of the adjustment indicated by the evident movement along joints probably took place before the structural deformation of the high erosion surface. Thus, exposures in the Unity mine show that such displacements commonly do not cut the andesite dikes that are correlated with the Columbia River lava, which elsewhere in north-central Idaho is involved in the structural movements resulting in the basins. It is possible, however, that some of the depression of the Warren basin was accomplished by movements too small to be reflected in the present topography along joints.

In the deepest part of the basin in the vicinity of Warren Meadows, normal faulting and warping have together resulted in a subsidence of approximately 2,000 feet below the general level of the surrounding divide.

As the high erosion surface was being elevated and tilted, and the basin was being relatively depressed, a drainage system was formed down the structural slopes into the basin.

The interior became an accumulating ground for material brought in through the system. The variety of unconsolidated sediments deposited in the basin, which includes unsorted material, presumably landslide and talus, beds of poorly-rounded boulders, well-rounded gravel, sand, clay, and peat, indicate that conditions were different at different times during the spasmodic settling of the basin.

At times, when the outlet was well established and had cut deeply below the level of its recent spillway, large volumes of sediments were removed from the basin. The streams of the interior drainage system correspondingly cut, particularly near their mouths, through the previously deposited sediments and into the rock floor of the basin as defined by the warped old erosion surface. Additional impetus to the canyon cutting along the lower parts of the interior-draining streams was given by the downfaulting, relative to the rest of the basin, of the central part that now constitutes Warren Meadows. Stream-cutting below the rock floor of the basin, resulting in the moderately steep valleys, was thus caused both by downcutting of the outlet level and by local downfaulting in the interior of the basin.

During the times that downwarping and downfaulting more than kept pace with the lowering of the basin's outlet accumulation of sediments in the interior was the dominant process, although the streams that brought in the material were, of course, eroding above the level of deposition in the interior. The sediments indicate that much of the deposition was on land, although some of the finer-grained material was probably laid down in a lake, or lakes, ponded behind the rising outlet.

These alternating conditions of removal and accumulation resulted in the formation of the gravel terraces around the Warren Meadows and along the lower parts of the interior streams and in the reworking of much of the material deposited in the basin early in its history.

At the present time, the deep canyons cut by the Salmon River and Warren Creek in the recent cycle as a result

> of the uplift and tilt of the old erosion surface have worked headward far enough to tap some of the interior of the basin through Houston, Shissler, and a few smaller creeks, and the Warren Creek Canyon that falls more than 3,600 feet in 10 miles is at the very threshold of the Warren Meadows and it will rapidly tap and dissect the remaining parts of the basin in a comparatively short span of geologic time." (Reed, 1937)

A similar description of the Secesh Basin (near Burgdorf) was presented by Capps (1943). The main difference in the descriptions is that Capps did not observe any Tertiary dikes in the Secesh Basin area. Larsen and Schmidt's (1958) road log through the subsection records quartz diorite on the west which changes to the typical coarse quartz monzonite (Ki3ch) in the central and eastern parts.

Subsection 3 - Chamberlain Basin Subsection

Subsection 3 is a large area of Idaho batholith bordered to the north by the Salmon River and to the south by a complex of Precambrian metasediments which are mapped as undifferentiated gneisses and schists (mpCl). Cater et. al., (1973) mapped most of the area as undifferentiated Idaho batholith which includes quartz diorite, granodiorite, and quartz monzonite. A zone of gneissic, leucocratic, quartz monzonite (Ki3cf) covers the northern and northeastern portions of the subsection. Cater describes the rock type as being older and representing greater depths than the adjoining quartz monzonite/granodiorite mass of the batholith. Contacts between the gneissic quartz monzonite and the batholith are sharp and in places crosscutting.

The linear map of the Elk City sheet shows a general linear trend striking north-northeast in Subsection 3 which is different from areas to the west (trend north-south) or to the east (general trend northwest). The general linear trends in the Elk City area are discussed at some length later (Section IV, general discussion).

Subsection 4 - Salmon River Canyon Lands Subsection

Subsection 4 is located on the Grangeville and Elk City AMS maps and includes the immediate drainage of the Salmon River including the South Fork and the Middle Fork of the Salmon and the Little Salmon River. Because it is a river drainage, Subsection 4 has little to do with lithology and more to do with drainage patterns.

Anderson (1934) discussed the Late Tertiary drainage changes in the Boise Basin area (see Section I, Subsection 10). Basically he described the effects of the Snake River downwarp and the compensatory upwarp to the north. In 1947, Anderson discussed the reasons for the drainage change in the Salmon River where it flows northward to Salmon, Idaho, and then abruptly shifts westward as it flows to Riggins. This paper is also of interest for a general discussion of other major drainages in the state and a portion of the work follows.

History of Drainage Development

"According to Ross, much of the western part of the area (site of the present Idaho batholith) has been a positive area since about the end of the pre-Cambrian and much of the eastern, a part of the Paleozoic geosyncline that extended alongside. When the region became involved in the late Jurassic orogeny and the intrusion of the Idaho batholith (Cretaceous), it all became land and supplied erosional debris to the neighboring Laramide trough. Thus, a northeastward drainage system apparently had become established even prior to the close of the Mesozoic era. The drainage apparently was maintained through the Laramide orogeny at the close of the Cretacous period, and, as indicated by the distribution of belts of Challis volcanics (late Oligocene or early Miocene) along old northeastward-extending valleys, well into Tertiary time.

The more immediate ancestral drainage probably played an important role in post-Challis planation, and the present northeast drainage pattern may be an inheritance from late Tertiary time. The evidence afforded by headwater streams and by present wind gaps indicates that the drainage was northeast and that the Continental Divide at that time coincided closely with the watershed of the present headwater streams. Except at its very head, where there have been some subsequent adjustments to faulting, Salmon River pursued its present northeast course; but instead of making an elbow turn a short distance below the town of Salmon, it continued on in a northeasterly direction, probably along the present inverted course of Carmen Creek and on toward the present Missouri River. The Salmon was then joined by its East Fork and also by Big Lost River. The latter flowed northeast through the site of Double Springs Pass and entered the Salmon not far below the present mouth of Pahsimeroi River by way of the wind gap across the northwest corner of Lemhi Range. Antelope Creek also flowed in a northeasterly direction through what is now Wet Creek Pass and through the unnamed pass at the head of Little Lost River in the Lemhi Range, but whether it eventually entered the Salmon or continued on as an independent drainage system has not been determined. A stream no longer in existence but which may be referred to as the "Arco River" then flowed northeast through Arco Pass and Pass Creek Pass and probably on across the present Continental Divide. At that time Chamberlain Creek, Middle Fork, and Panther Creek had no local connection with Salmon River but may have joined it some miles beyond the present state line. Thus, the drainage just prior to the crustal disturbance

that came at the close of the Tertiary was an integral part of a northeast drainage system that extended on into Montana and presumably to the Arctic or the Gulf of Mexico.

Disruption of this northeast drainage system probably started with the orogeny that came at the close of the Tertiary period. The general uplift apparently favored the vigorous headward erosion of a stream from the west which captured the now barbed tributaries, Chamberlain Creek and Middle Fork, then Panther Creek, and, finally, the main Salmon River. At or about the same time, rising block mountains south of the Salmon River barred the way of other northeast streams, cutting off and diverting one of its main tributaries, Big Lost River, southeast to the Snake River Plain. Antelope Creek also was forced to abandon its northeast course and to become a tributary of Big Lost River. Consequent streams developed within the structural basins, fed by streams from the bordering mountains. Warm Spring Creek, Pahsimeroi River, and Lemhi River flowed northwest to join the Salmon. Little Lost River and Birch Creek drained southeast along the structural basins to disappear on reaching the Snake River Plain.

More or less closely associated with the mountain uplift was the Snake River downwarp. While the mountains and intervening basins were being arched along an axis parallel to and about 50 miles from the margin of the Snake River Plain, the Snake River downwarp was carrying the ends of the ranges to the level of the plain and then below, where they are concealed beneath a covering of lava that partly filled the great downwarped basin. By the downwarp, the headwaters of Arco River were brought below and became a part of the underfloor of the Snake River Plain and were completely lost from view, with only Arco and Pass Creek passes left as evidence of their former existence.

Since the captures and diversions, the streams have become established in their new courses and are continuing to deepen their valleys in the mountainous areas and to aggrade their floors where they occupy the structural basins."

Conclusions

"Piracy and large-scale faulting have had an important role in disrupting a northeast Tertiary drainage system and in fixing the pattern of existing stream systems in the northern Rocky Mountains of east-central Idaho. Piracy has virtually reversed the direction of flow of

> Salmon River and turned its waters from the Missouri system to the Snake and Columbia rivers and the Pacific Ocean. Block faulting has also diverted some of the tributaries of Salmon River and other streams into structural basins extending southeast to the Snake River Plain. Downwarping involved in the origin of the Snake River Plain has caused the bordering mountains to plunge and disappear beneath the great lava-filled depression and to carry from sight the headwaters of streams that formerly drained to the northeast. As a consequence of these diversions the Continental Divide was shifted about 100 miles east of its location in late Tertiary time." (Anderson, 1947)

There are two types of rocks over which the Salmon flows, high grade metamorphics to the east and west, and the Idaho batholith complex in between. The rocks on the Elk City map are discussed in more detail elsewhere while a description of the rocks within the Salmon drainage near Riggins (Grangeville AMS map) follows.

Hamilton (1963), described the rocks in the Riggins area in detail. The part of Subsection 4 on the Grangeville map neatly outlines most of the metamorphic complex called the Riggins Group. The Riggins Group consists of four major units which are from youngest to oldest: the Squaw Creek schist (6,000 feet thick, m24), the Berg Creek amphibolite (1,200 feet thick, m23), the Lightening Creek schist (8,000 feet thick, m22), and the Fiddle Creek schist (9,000 feet thick, m21). These rocks are thrust onto the Seven Devils volcanics to the west. Hamilton's abstract is included here as it gives a good overall picture of the lithologies and structure in the Riggins area.

Metamorphism in the Riggins Region, Western Idaho

Abstract

"The west-directed Rapid River thrust separates two assemblages of metamorphic rocks in the northwestern part of the 30-minute Riggins quadrangle. Below and west of the fault is the metavolcanic formation of the Seven Devils Volcanics (Permian and Triassic), the schistose marble and crystalline limestone of the Martin Bridge Limestone (Triassic), and the Lucile Slate (Triassic (?)). These rocks are of low metamorphic grade throughout; granitic rocks intrusive into them have been metamorphosed in the same degree. The thick Seven Devils Volcanics consist of submarine lavas, tuffs, and agglomerates with the compositions of basalt, andesite, dacite, and their albitized equivalents spilite, keratophyre, and quartz keratophyre. The Martin Bridge Limestone was approximately 2,000 feet thick prior to deformation. The Lucile Slate is thin.

Lying tectonically above and east of the Permian and Triassic sequence is a thick sequence of metamorphic rocks, dominantly schists, here named the Riggins Group. The present order--which is probably in part tectonic rather than stratigraphic--within the group is, from bottom to top, Fiddle Creek Schist (metamorphosed silicic tuffs and lavas), now about 9,000 feet thick; Lightning Creek Schist (metamorphosed mafic to silicic tuffs and lavas), about 8,000 feet thick; and Squaw Creek Schist (metamorphosed sedimentary and volcanic-sedimentary rocks), about 6,000 feet thick. Metaperidotite is intercalated at or near the contact between the Squaw Creek and Lightning Creek Schists, and this contact is probably a premetamorphic thrust fault. An amphibolite unit lies between the same schists in another area.

Chemical analyses are presented for major and minor elements in 30 specimens, mostly metavolcanic rocks of the Riggins Group. Volcanic compositions vary from calc-alkaline to sodic and define a compositional spectrum comprising normal to keratophyric rocks. The metavolcanic rocks are compositionally identical with the older and variably spilitized volcanic rocks of the Aleutian Islands, where volcanism has continued to the present time, and where subaerial rocks are of normal calc-alkaline compositions. In both the Riggins quadrangle and the Aleutian Islands, exchange of magmatic calcium for seawater sodium probably resulted in the sodic rocks.

The rocks of the quadrangle increase in metamorphic grade eastward and southward. Some of the common metamorphic index minerals of pelitic schists are lacking; and isograds have been drawn, in order of increasing metamorphism, upon the westernmost appearance of aluminian prochlorite (as opposed to the ferroan variety), biotite, clinozoisite (as opposed to epidote), garnet, oligoclase, and andesine. The metamorphism of the Riggins Group probably occurred during Early Cretaceous time, whereas the rocks beneath the Rapid River overthrust may have been metamorphosed not only at this time, but also during the Late Jurassic.

Analysis of data from this area and many others in which progressive metamorphism has been studied indicates that the heat represented must have been partly introduced by magmas mobilized at deeper levels and partly generated by deformation of the rocks. The generally accepted theory that the heat of regional metamorphism is due to depth of burial seems largely invalid." (Hamilton 1963)

Hamilton noted that the metamorphic rocks of the Riggins Group are essentially confined to the Riggins Quadrangle and that they form a lens shaped pod in this area. Hamilton's map (1969) contains an excellent fence diagram which shows the complex structural relationships which exist in the quadrangle.

Subsection 5 - Big Horn Crags Subsection

The northern two thirds of Subsection 5 is underlain by the Crags pluton which is Tertiary in age (dated at 44.2 - 47.5 m.y.). Cater et. al., (1973) described the pluton as a medium to coarse grained, porphyritic, biotite-quartz monzonite which grades to granodiorite. High angle jointing has produced spiral peaks and steep clifts. Large exfoliation knobs surrounded by deep grus are observed in the higher elevations. These areas were obviously not scoured by Alpine glaciation which greatly affected lower elevations. The southern portion of the subsection is part of the Precambrian metamorphic complex consisting of the Yellowjacket and Hoodoo Formations which were discussed in Section I, Subsection 23.

Subsection 6 - Panther Creek Canyon Lands Subsection

The only work which has been done near Subsection 6 is Anderson's (1943) report on the Cobalt area which is discussed in Subsection 7.

Subsection 7 - Panther Creek Basin Subsection

A major part of Subsection 7 is covered by the Leesburg quadrangle which was mapped by Shockey (1957). The subsection is drained by Panther Creek or smaller tributaries of the Salmon river. Rock units in the area consist of + 35,000 feet of Precambrian rocks, the Kinnikinnic quartzite (± 4,000 feet), upper Challis pyroclastic and sedimentary units and some Idaho batholith intrusives.

The Precambrian lithologies are described by Shockey as follows: ± 10,000 feet of a formation which consists of gray phyllite, graywackes quartzites, and black shale units, and a formation consisting of ± 25,000 feet of impure gray quartzite with purple quartzite and red slate in the upper part. The first Formation is the same as the Apple Creek phyllite (mpCs3) described by Anderson (1956) just to the east of the Leesburg quad and the thick quartzite is probably equivalent to the Swauger/Lemhi quartzites which overlie the Apple Creek phyllite. This quartzite is designated on the map as mpCsud, for metamorphosed Precambrian sediment undifferentiated as Shocky did not map Swauger or Lemhi equivalents separately.

The Kinnikinnic quartzite has been described earlier in the introduction to Section III, as have the upper volcanoclastic members of the Challis volcanics (i.e., Kirtley Formation, Tmud). Shockey mapped two sets of glacial deposits which he thought were Wisconsin and Pre- Wisconsin.

Major structural features include the Leesburg anticlinorium which strikes N 45° W through the central part of the quad. The anticlinorium forms in part, the northern portion of the Lemhi range. The anticlinorium is paralled by the Leesburg fault which is a high angle reverse fault with at least 17,000 feet of displacement. Most of the structure in the area could have been produced by tectonic forces directed S 45° W and were probably associated with the Late Mesozoic/Early Tertiary Laramide orogeny. Shockey noted that apophyses of Idaho batholith intrude the quartzites which are batholith roof rocks in this area.

Ruppel (1973) has discussed the relationship of the Yellowjacket Formation west of the Leesburg quad, and the Apple Creek phyllite/ Swauger quartzite to the Belt rocks of northern Idaho. The comparison follows as it ties all of these rock types into one framework.

Precambrian Sedimentary Rocks in East-Central Idaho

Abstract

"Precambrian sedimentary rocks in east-central Idaho crop out in a belt that reaches 150 miles from the Snake River Plain northward to the Idaho-Montana line at Lost Trail Pass. In the northern part of this belt, the rocks are mostly fine-grained, dark-colored, micaceous, feldspathic quartzites that closely resemble the Yellowjacket Formation. Farther south, the Yellowjacket-like rocks are overlain by a thick sequence of lighter colored, fine-grained feldspathic quartzites and argillitic siltites that can be separated into a number of redefined and new mappable units, with a unit of pink medium-grained non-feldspathic quartzite at the top. The thickness of the Yellowjacket is not known, but probably exceeds 10,000 feet; the overlying rocks are at least 30,000 feet thick.

The eastern boundary of these Precambrian sedimentary rocks appears to be tectonic, and regional considerations suggest that at least the sequence above the Yellowjacket has been moved from its original depositional area eastward perhaps as much as 100 miles on major flat surfaces of detachment.

Correlation of the east-central Idaho Precambrian rocks with the better known Belt Supergroup in western Montana and northern Idaho remains tenuous because of the differing characteristics of the sequences and the likelihood that they were deposited under different conditions--the Belt Supergroup in the Belt Basin, and the east-central Idaho rocks in the Belt miogeosyncline. However, possible correlations are suggested by a few events and lithologic features that seem to be common to both sequences. The Yellowjacket and Prichard

> Formations are both basal sequences that differ greatly from the overlying rocks; they are possible correlatives. The Apple Creek Formation of A. L. Anderson, in the middle of the sequence above the Yellowjacket, is the only carbonate-bearing formation in east-central Idaho, and perhaps is equivalent to the middle-Belt carbonate rock unit represented by the Wallace Formation and Helena Dolomite; the rocks between the Yellowjacket and Apple Creek thus occupy a stratigraphic position similar to that of the Ravalli Group. The rocks at the top of the east-central Idaho sequence, pink medium-grained non-feldspathic quartzites, were named the Swauger Quartzite by C. P. Ross. These rocks differ so strikingly from the rocks below as to suggest a fundamental change in sedimentary environment, perhaps reflecting the tectonic adjustments that took place during Missoula Group deposition; the rocks between the Apple Creek and Swauger therefore could be equivalent to part of the Wallace Formation or to part of the Missoula Group, and the Swauger could be equivalent to part or all of the Missoula Group." (Ruppel)

Anderson (1943) described the area around Cobalt, Idaho, which is near the junction of Subsections 5, 6, and 7. He did not map the area but noted that is was an area of metaquartzites and schists which increased in metamorphic grade from south to north towards the edge of the batholith. He described an outlier of the batholith on lower Panther Creek and noted that the major structure in the area appeared to be a broad fold (ed. note: This fold is probably parallel to the northwest trending folds in the Leesburg quad.).

Subsection 8 - North Fork Salmon Breaks Subsection

No geologic information is available.

Subsection 9 - West Fork Bitterroot Streamcut Subsection

Subsection 9 is discussed under Section V, Bitterroot Mountain Section.

Subsection 10 - State Divide Uplands Subsection

(See Winegar, R. C., Bitterroot II, Geologic Survey. February 1974. A special Forest Service Contract Report.)

Subsection 11 - Devils Point Weakly Glaciated Lands Subsection

The southern half of Subsection 11 was mapped by Weiss et. al., (1972) and the northern half was reconnaissance mapped by Greenwood and Morrison (1973). The subsection contains the headwaters of the Selway River. The Idaho batholith rocks in the area were mapped as quartz monzonite (Ki3c) by Greenwood and Morrison and as granodiorite by Wiess et. al. This is typical of the overlap in classification of the rocks in the batholith. A few outcrops of garnetiferous biotite quartzites and pelitic schists and gneisses (mpC12) outcrop in the northern portion of the area and are tenatively correlated with the

Ravalli Group of the Belt Supergroup (Greenwood and Morrison, 1973).

Some small outcrops of augen gneiss (m36) were mapped in the area. Many outcrops of augen gneiss occur on the Elk City map and these seem to be in a diffuse band which conforms to the northwest tectonic trend in the region. This band runs from the southeast corner of the Elk City map northwest to Elk City.

A portion of the southeast corner of Subsection 11 was mapped as foliated quartz monzonite (Ki3cf) by Greenwood and Morrison and as undifferentiated gneisses by Weiss et. al., which is characteristic of this area. The geology is extremely complex and not well understood.

Weiss et. al., (1972), described glaciation in the Salmon River Breaks area a bit differently from other observers. This discussion follows.

Salmon River Breaks Primitive Area, Idaho

Glaciation

"Ridges higher then 7,000 to 7,200 feet had permanent snowfields on their north and northwest sides during the late Pleistocene, and ridges higher than 8,000 feet supported active glaciers. Most south- and southwest-facing ridges show no evidence of glaciation below 8,000 feet and little above that elevation.

Glaciers from higher centers of accumulation flowed mostly eastward down preexisting valleys, carving them to well-developed U-shapes down to elevation of about 6,000 feet. The lowest recognized ice terminus was at about 5,000 feet in the valley of Ring Creek, a tributary of Sabe Creek.

Moraines are small and thin and occur mostly as remnants of lateral moraines in the lower parts of a few valleys. Best examples are along streams that drain the east side of Stripe Mountain. Terminal and recessional moraines are almost entirely lacking. Because the moraines are small and of no economic value, they were not mapped. Cirques are moderately steep walled, but only a few on the highest peaks are major obstacles to foot travel. Many cirques contain lakes which, so far as is known, are all in rock-enclosed basins."

Studies Related to Wilderness-Primitive Areas

"All glacial features in the study area are fresh. Little or no weathering has occurred on glaciated surfaces, moraine remnants still support undrained depressions, and many undrained depressions contain water rather than silt or vegetation. Furthermore, no

evidence for multiple stages, or even for minor readvances during recession, was noted. This apparently youthful, simple glacial history is in contrast to other parts of central Idaho (Ross, 1930).

The absence of older glacial features may be the result of more extensive younger glaciation that eliminated the older deposits, or it may be because no earlier glaciation took place. Topography suggests a third possibility. The entire Salmon River Breaks is an area of deep, steep-walled canyons and relatively narrow ridges. In an area of such steep gradients, evidence of older glaciation may simply have been eroded away.

It may be possible to either confirm or correct the age suggested for existing alpine glacial deposits by a study of the gravel bars along parts of the Salmon River Canyon. In places tributaries dumped large quantities of gravel into the main stream, building bars that are locally 20 to 50 feet thick and as much as a quarter of a mile long. These bars may have formed during periods of increased flow from tributaries. Those periods may coincide with periods of glacier advance on the ridges. Freshness of the material in the bars suggests that the deposits are late Pleistocene in age. Since the period or periods of deposition, the Salmon River has been reexcavating its canyon and has cut through several of the bars to reexpose the old bedrock floor of the canyon. Such bars can be seen at the mouths of Rattlesnake Creek, Bargamin Creek, Bear Creek, and Big and Little Squaw Creeks." (Weiss, 1972)

Subsection 12 - Saba-Magruder Mountain Moderately Glaciated Uplands

Subsection 12 is drained by the tributaries of the Selway River. The northwestern border of the Subsection coincides with Bargamin Creek which flows to the Salmon. A major fault system extends along Bargamin Creek and goes far to the north (see Subsection 22) and southwest along the Salmon River. The fault is associated with several Tertiary dike swarms. Greenwood and Morrison (1973), described the Bargamin Creek-Burning Creek segment of the fault as being a zone of shearing and dense dike intrusives. Cater et. al., (1973) mentioned the Bargamin Creek fault further to the southwest and believed that some strike-slip movement had occured along the fault. The linear map of the Elk City region shows that a series of north trending linears go through this Subsection. Most of the major drainages are aligned on these linears and the whole area is probably a series of grabens and horsts with some strike-slip movement on the faults adding to the complexity.

Most of the subsection is underlain by Idaho batholith mapped as quartz monzonite or foliated quartz monzonite. The southwestern portion contains some Wallace equivalent schists (mpC13) and a few minor Tertiary intrusives were mapped in other parts of the region.

Subsection 13 - Red River Basin Uplands Glaciation Subsection

Subsection 13 is drained by Crooked Creek which flows south to the Salmon, and the Red River system which flows north to the American River which in turn flows west to the North Fork of the Clearwater River. The southern portion of the Subsection is partially drained by Big Creek which flows to the Salmon.

Much of the geology of 13 is similar to Subsection 19 to the north. The primary difference is that the grade of metamorphism in 13 is higher than in 19. Biotite schists (mpC11) which are regarded as probable Prichard equivalents in 19 outcrop over much of 13 (Reid, 1959). Large masses of augen gneiss (m36) are present in the northeastern part of 13 and a large outcrop of quartz monzonite (Ki3c) of the Idaho batholith is located just south of the augen gneiss complex.

The Elk City linear map shows that most of the major drainages are probably controlled by north-south striking faults which supports Reid's (1959) idea of valley development by faulting in this region (see Subsection 19). An important difference exists between the linear patterns in Subsections 13, 14, 18, and 19, and the patterns to the east and north of these subsections. The linears to the east trend north northeast while to the north the linears are associated with the Trans-Idaho discontinuity. Many mining areas are located in 13, 14, 18, and 19, and the different type of linear patterns may be related to the mineralization (see general discussion of Section IV for more details).

Subsection 14 - Buffalo Hump Glaciated Lands Subsection

Subsection 14 is drained by the tributaries of the South Fork of the Clearwater River to the north and by tributaries of the Salmon River to the south. Shenon and Reed (1934) mapped the eastern part of 14. They differentiated between quartzites and schists in the Precambrian metamorphics but these have been lumped as Prichard Formation equivalents (mpC_{11}) as they are probably the same units mapped just to the east by Reid (1958). The Idaho batholith was described as granodiorite/quartz monzonite in composition and locally these rocks graded to gneisses as did some of the quartzites. Many andesite dikes cut the area and the major structural trend is to the northwest but few other particulars about structures were mentioned. The development of the topography in the subsection is similar to the topographic history of the Newsome and Elk Creek Basins to the east (see Reid's 1959 discussion in Subsection 19).

Little information is available about the western portion of 14 but it is probably similar to Subsection 2 to the south and Subsection 15 to the north which are almost entirely Idaho batholith. Most of the major drainages and basins seem to coincide with north-south or northeast trending linears which are probably faults which have controlled basin drainage development.

Subsection 15 - Florence Basin Subsection

Subsection 15 is underlain by granodiorite/quartz monzonite of the Idaho batholith according to Hamilton (1963). The western edge of the subsection marks the boundary between batholith (Ki3b) and marginal gneissic quartz diorites (mKi2a). Hamilton noted that much of the western part of 15 is covered with glacial deposits.

Reed (1939) worked slightly to the east of Hamilton in the Florence mining area and noted that much of the area consisted of quartz diorite of the batholith with a few small bodies of quartzite and schist. The quartz diorite was hydrothermally altered and highly weathered and is representative of the border facies of the batholith. Reed noted that the batholith rocks changed in composition from quartz diorite near Florence to granodiorite and finally quartz monzonite going eastwards to Elk City. Reed described the schists and quartzites as small zenoliths or roof pendants which were intruded by pegmatites.

Reed mentioned evidence for two periods of glaciation and a later stage of fluvial deposition in the major valleys. He noted that Capps (pers. comm.) thought that beavers had moved into the area after the Wisconsin Ice Age and built dams which had caused the deposition of the fluvial valley deposits (characterized by peat bogs).

Reed discussed the history of the development of the drainage system in the Florence areas as follows.

> Tertiary Structure and Development of Topographic Features
>
> "One of the most striking physiographic features of central and north-central Idaho is the high plateau, now greatly dissected, that stretches all the way across the Clearwater and Salmon River mountains. This surface was described and pointed out that the altitude of the surface increases from about 7,000 feet in the northwest part of the Clearwater Mountains eastward to 9,000 feet or more in eastern Idaho.
>
> This surface is the so-called "summit", or Idaho peneplain, but it should be emphasized that, since it has a local relief of at least 500 feet within a mile, it is not a true peneplain. Its age, extent, and manner of development are somewhat controversial and have been discussed by several authors.
>
> This widespread, high surface was formed by stream erosion and reached its maximum development at a time when central Idaho lay at a much lower altitude than at present. This erosion cycle cut deeply into the batholith and its enclosing rocks. Coarse-grained rocks, which must have crystallized at a considerable

depth, crop out extensively on it. Erosion, however, did not pare deeply enough into the rocks to destroy the batholithic roof completely as is witnessed by the presence of large masses of older rocks, some of which are presumably roof pendants which are still preserved on the high surface.

Obviously, the development of the surface was completed after the consolidation of the batholith, which probably took place in Upper Cretaceous time. There is evidence that erosion of this surface continued until after lower Miocene time. Near Thunder Mountain, in the Idaho National Forest south of the Salmon River, the surface truncates rocks of the Challis volcanics, which are probably of upper Oligocene or lower Miocene age.

Little is known of the Tertiary drainage system on this old erosion surface. The relatively gentle relief indicates that the whole area did not stand much above base level. The predominant northward trend of the rocks in central Idaho leads to the inference that the stream courses and the divides in general probably trended north or south. The stream valleys must have been broad and the gradients gentle, and no doubt contained considerable deposits of detrital material.

A lower erosion surface, less widespread than the one just described, has been recognized by Capps and Reed in the Nezperce National Forest, and probably is present also in other areas. The lower surface lies in general between 500 and 1,500 feet below the higher one. Capps has recognized that the lower surface is well developed in an east-west belt across the Buffalo Hump quadrangle roughly parallel to, and on both sides of, the South Fork of the Clearwater River.

As recorded by the presence and pattern of the lower erosion surface, the sequence of events appears to be roughly as follows: After the formation of the high surface, and while it still lay near base level, the whole area began to be uplifted. The east and southeast parts were raised higher than the rest, thus imparting a general northwesterly tilt to the old surface. Tributary streams flowing west had their gradients steepened. Those flowing east had their gradients lowered. The northward-trending divides were lowered in places by the steepened westward-flowing streams. Possibly some ponding took place in the broad northward or southward-trending valleys so that overflow from the east across some of the lowered divides took place. Thus the major streams found courses toward the northwest, initiated and controlled by the northwestward tilt of the old surface, and the present courses of such rivers as the South Fork of the Clearwater and the Salmon

were determined. With continued uplift, the erosive power of these major streams became stronger and they began to lower and broaden their valleys from downstream headward. A sufficiently long pause in the uplift allowed erosion to go far enough to develop the lower surface nearly to base level, but not far enough to remove large remnants of the higher surface. The areas east and southeast of the Buffalo Hump quadrangle appear to have lain so near the heads of the streams that the lower surface was not well developed in them. This may account for the fact that the lower surface has not been recognized elsewhere.

The Florence surface is believed to be part of the lower erosion surface. The suggestion is offered here that the Salmon River once turned north at about the present location of Florence and jointed the South Fork of the Clearwater near Castle Creek Ranger Station, and so developed the Florence surface. The present northward-flowing part of Salmon River from Riggins to below Whitebird might develop just such a surface if a local base level below Whitebird were present for a long enough time.

Columbia River lava is present on the Florence surface at several places, such as Hungry Ridge, near McConnell Ranch on Mill Creek, in the vicinity of Adams Ranger Station, on Royal Creek in the Bungalow district, and elsewhere. Although later disturbed, as discussed below, the flows that spread out on the Florence surface must have been very nearly the highest, youngest flows of the great lava sequence. But at several places Latah sediments, identified by good Latah floras, are interbedded with these flows and fix their age as middle Miocene. Thus, the younger surface was developed before the middle Miocene.

After the development of the younger erosion surface, central Idaho was again uplifted in a series of repetitive steps to its present altitude. The base level of erosion was thereby greatly lowered, the rejuvenated streams began to cut actively, and the present system of deep and steep canyons, as impressive as any in the United States, was formed. The streams now flow in the canyons far below even the lower erosion surface, and are still cutting rapidly.

During this uplift, the crustal block that comprises central Idaho was broken in many places by long normal faults of considerable displacement. Crustal warping probably took place also. The normal faults without exception, so far as is known, trend northward about parallel to the strike of the bedrocks. In most of them, the east side is the downthrown side.

The faulting and warping resulted in the formation of the many basins within the mountains from which has come much of central Idaho's placer gold.

More of the normal faults are known near the western border of the mountains against the Columbia Plateau than elsewhere. They are generally recognized by the scarps that are still conspicuous features of the landscape and by the observable displacement of the younger erosion surface, as marked by the base of the Columbia River lava flows. The faults may be more numerous far within the mountains than is generally recognized, because the lava flows are not present there and the faults are much more difficult to identify.

The Columbia Plateau itself has been depressed as a unit relative to the crustal block embracing the mountains. How much of the observable displacement is due to actual depression of the plateau, and how much to the elevation of the mountains, is not known. This major displacement of the two large crustal units in the Nezperce country amounted to 2,000 or 3,000 feet and was accomplished by normal faulting and by steep monoclinal warping. The west side is the downthrown side in the displacement of the plateau and mountain blocks.

The essential points of evidence of the displacement are as follows:

(1) The lava surface of the plateau is everywhere essentially parallel to the flows and the plateau has probably not been lowered greatly by erosion.

(2) The general altitude of the Columbia Plateau against the Clearwater Mountains is between 3,000 feet and 4,000 feet.

(3) Many remnants of lava on the lower erosion surface near the border of the mountains lie at an altitude of 6,000 feet or more.

(4) Faults and monoclinal folds showing the displacement can be identified along the mountain front.

(5) Wherever observed the basal lava flows appear to have poured out on a land surface. The lava sequence of the Columbia Plateau has now been depressed so that over large areas at least 1,000 feet of the sequence is below sea level.

The Florence surface is one of the best examples known of a depressed part of the mountain block. Normal faults can be traced along its western edge almost continuously from north of the South Fork of the Clearwater into the

Salmon River Canyon, a distance of more than 30 miles, by means of their scarps and by downfaulted lava remnants. At some places, the total displacement, which is as much as 1,000 feet, appears to be along one fault. At other places, at least two faults are present.

There is no apparent fault on the east side of the Florence surface, and there the relatively steep rise to the higher erosion surface may be due entirely to the erosional break between the two surfaces or it may be accentuated by warping. It appears significant that the eastern edge of the Florence surface is relatively sinuous and that the surface makes broad indentations eastward up both the Salmon and the South Fork of the Clearwater.

The displacement of Columbia River lava flows along the faults at the western edge of the Florence surface must have followed the consolidation of some of the highest of the flows. The movements apparently ceased before a pre-Wisconsin glacial stage. Much of the eastern part of the Florence surface is deeply covered by a mantle of detrital material that appears to be in part at least of glacial origin and this mantle is not disturbed.

The pre-Wisconsin glaciation took place after the uplift of the mountain block and before the youthful cycle of erosion, which was caused by the uplift and which resulted in the present deep canyons, for the mantle of detritus on the eastern part of the Florence surface bears no relation to the present canyons that dissect the surface.

The Wisconsin glaciers, on the other hand, pushed down the valleys carved in the Florence surface and their moraines are found in some of them hundreds of feet below the surface, for example, in the canyon of the main eastward tributary of Slate Creek south of Adams Ranger Station." (Reed, 1939)

Subsection 16 - John Day Subsection

Subsection 16 is underlain mostly by an igneous rock called Trondhjemite (Ki3T). Trondhjemite is an oligoclase-biotite quartz diorite which has virtually no potassium feldspar and is only found near the border of the batholith. The northeastern part of 16 is covered by the Martin Bridge limestone (mTrc) which is Triassic and overlies the Seven Devils' volcanics. The limestone is in turn overlain by the Lucille slate (mTr3) which outcrops to the west in Subsection 4. The limestone is sheared, recrystallized, and very deformed. Subsection 16 is drained by tributaries of the Rapid River and is separated by 15 to the east by a fault.

Subsection 17 - Dairy Mountain Dissected Plateau Lands Subsection

The major feature of Subsection 17 is the Columbia River basalt which covers most of the area. Seventeen is drained by tributaries to the Salmon River. Bond described the Columbia River basalt in his paper on the Clearwater Embayment (1963). The Clearwater Embayment is a plateau (elevation, 2,500 - 4,000 feet) characterized by low, undulating hills on its surface. Certain parts of the plateau are covered by loess deposits called the Palouse Formation (Qw) and in some places thin beds of recent ash falls are observed.

The Columbia River basalts can be subdivided into a lower basalt (TelCl) which correlates with the Picture Gorge flows in Washington and Oregon and an upper basalt (TelCu) which correlates with the Yakima basalt in Washington and Oregon. Bond's field descriptions of the two basalts follows:

Field Identification of Formations

Hand Specimen Appearance

"Although the occurrence of phenocrysts in "Lower" Basalt flows makes recognition of the two formations easy, this criterion may not be infallible. Russell (1901, p. 31) reported some flows beneath the porphyritic basalt which are "normal and in places markedly characteristic" (non-porphyritic) in the Snake River canyon. Within the Clearwater Embayment, however, every flow of "Lower" Basalt has labradorite phenocrysts up to one-eighth of an inch thick which range from one-fourth to one and one-half inches in diameter. The low density of distribution in some flows may necessitate the taking of several hand specimens to expose a large phenocryst.

With the exception of samples from the top flow of "Upper" Basalt (Lolo Creek Flow) less than one hand specimen in one hundred displays a large phenocryst. The uppermost flow contains scattered crystals of plagioclase up to three-eights of an inch in diameter; the phenocrysts most typically are from one-sixteenth to one-fourth of an inch across.

The presence of large phenocrysts is not the only lithologic difference between the two formations. As indicated by Waters (1961, p. 591), minor differences in the mineralogy of basalt flows may result in differences in their physical appearances. The following lithologic comparisons can be made between "Lower" and "Upper" Basalt flows in the Clearwater Embayment.

The color of unweathered "Lower" Basalt is commonly a dark brown with some fresh surfaces of the more dense

portions of the flows showing deep purplish to sooty tints. A waxy luster is typical. The phenocrysts on fresh surfaces are clear and appear the same color as the groundmass. They exhibit a resinous luster which might best be compared to that displayed by large cleavage surfaces of sphalerite. Waters (1961, p. 599) discussing the Picture Gorge Formation in the Imnaha River canyon, credits slight deuteric alteration for this "greasy" appearance of the rock.

"Upper" Basalt, on the other hand, is blue-black in dense rock and charcoal to slightly lighter shades of gray in the less slowly cooled segments of flows. Minute idiomorphic crystals give some surfaces a felty texture.

On weathered surfaces the chocolate color of the groundmass of "Lower" Basalt flows is lightened to brown or orange-brown. A speckled or mottled appearance is created by the phenocrysts which weather to a contrasting light brown or tan.

Weathered "Upper" Basalt is very light gray to light blue; the more complete the alteration to clay, the lighter is the color.

Chlorophaeite, permeating some of the basalt of both formations as void fillings, may tint freshly broken surfaces green or blue-green. Because the mineraloid is very susceptible to alteration, only portions of the more dense massive flows tend to preserve this characteristic.

Both "Lower" and "Upper" Basalt flows are finely crystalline; a grain size less than one-third of a millimeter is common for the bulk of the groundmass. The size and percentage of vesicles varies somewhat inversely to the distance from the top and bottom contacts of individual flows. Microvesicular or "diktytaxitic" openings (Fuller, 1931b, p. 116), as qualified by Mackin (1961, p. 6), are present away from chilled and quenched zones particularly in "Lower" Basalt flows and in the highest flows of "Upper" Basalt. In such weathered "Upper" Basalt flows the minute angular voids may be filled with yellow clay.

Outcrop Appearance

"Lower" Basalt flows consistently show basal colonnades which compose up to thirty percent of the total flow thickness. The colonnades are characterized by very well developed columnar jointing; the columns range from less than one and one-half feet to more than six feet in

diameter. Although typically vertical, the columns may be curved or inclined due to uneven cooling caused by irregularities in the surface over which the flow spread. Slightly more resistant to erosion than the entablatures of the flows, the larger and more perfectly formed vertical columns of the colonnades of a few of the thicker flows stand out like rows of vertical pillars beneath smooth and rounded benches which slope gently canyonward.

Wavy and irregular jointing of the entablatures is very well developed in the centers of "Lower" Basalt flows and typically gives way upward to vertical columns which comprise about the upper ten percent of most flows. The continuity of individual columns in the entablatures of "Lower" Basalt flows remains very prominent and distinct even where they are in sets pronouncedly curved or inclined to nearly horizontal. In some flows hexagonal jointing is so regular that segments, one to two feet long, have been used as building blocks. Fan-jointed "rosettes" are typical of some flows of "Lower" Basalt and were observed by Shenon and Reed (1936, p. 134) to exceed 50 feet in breadth along the bottom of the Salmon River canyon.

In "Upper" Basalt the jointing is much less systematic and successive flows, although individually consistent in joint characteristics, vary from flows with well developed entablatures and colonnades to flows displaying no such recognizable subdivisions. If present, although not completely without exception, the uniform columns of the colonnade are restricted to within a very few feet of the base of the flow.

The entablatures of the thicker flows of "Upper" Basalt, particularly near the bottom of the formation in the Salmon River area, have as many as six tiers of vertical columns. The columns are from four to ten inches in diameter. The jointing of the entablatures of the thinner flows is commonly of a "brick-bat" type. Fan-jointed or curved columns are present in "Upper" Basalt flows also, but the persistence of any one column through a rosette is seldom observed. The columns tend to bifurcate and have twisted appearances and scalloped joint surfaces.

Under prolonged exposure "Lower" Basalt flows deteriorate into a coarse granular sand composed of fragments of phenocrysts and groundmass. The detritus is similar in color to the weathering parent material. This type of granular disintegration, characteristic of some flows in the Steens Mountains of southern Oregon, was ascribed

by Fuller (1930, pp. 221-227) to endomorphic alteration of the solidifying basalt. He noted that alteration is greater away from both the rapidly chilled contacts of flow tops and bottoms and the joint surfaces of individual columns. The alteration probably represented the corrosive effect of volatiles trapped within a column's interior after congealing of the surface of a flow and the formation of internal joint surfaces. The mechanics of the disintegration was credited by Waters (1961, p. 599), in Picture Gorge Basalt, to swelling of deuterically altered olivine and chlorophaeite when subjected to weathering processes.

Another characteristic of "Lower" Basalt mentioned by Waters, which results from the expansion of clay minerals, is a marked impermeability caused by the alteration of flow centers and the swelling shut of columnar joints within the colonnade of the basalt flows. He found that the line of springs consequently created at the top of the formation was useful for stratigraphic purposes. The presence of a number of spring-lines within the canyons of the Clearwater Embayment, due in part to a higher average rainfall plus a larger ground-water supply from marginal and steptoe catchment areas, makes springs less useable for formation identification but satisfactory for tracing flow contacts.

Exfoliation structures are far more pronounced in "Lower" Basalt flows and are another expression of the tendency of the rock of this formation to swell markedly by the expansion of clay minerals. Exfoliation has been very effective in rounding joint edges of exposed columns.

The sharp-edged, solid-appearing columns of "Upper" Basalt flows, less susceptible to internal disintegration and decomposition, part mechanically along joint and fracture planes and create great piles of relatively unweathered basalt talus at the base of slopes. These accumulations tend to bury underlying flows so completely in some canyons that the presence of "Lower" Basalt may be difficult to establish.

Two distinct types of residual soil are recognized in the canyons of the Snake and Salmon Rivers by the Soil Conservation Service. The Tripod Rocky Loam is restricted to the lower portions of the canyon walls at elevations beneath the Gwin Silt Loam. The properties which make these two soils distinct have been inherited in part from their parent basalt material, the "Lower" and "Upper" Basalts respectively. Useful characteristics for stratigraphic identification in reconnaissance of the two formations are the higher water-holding capacity, greater

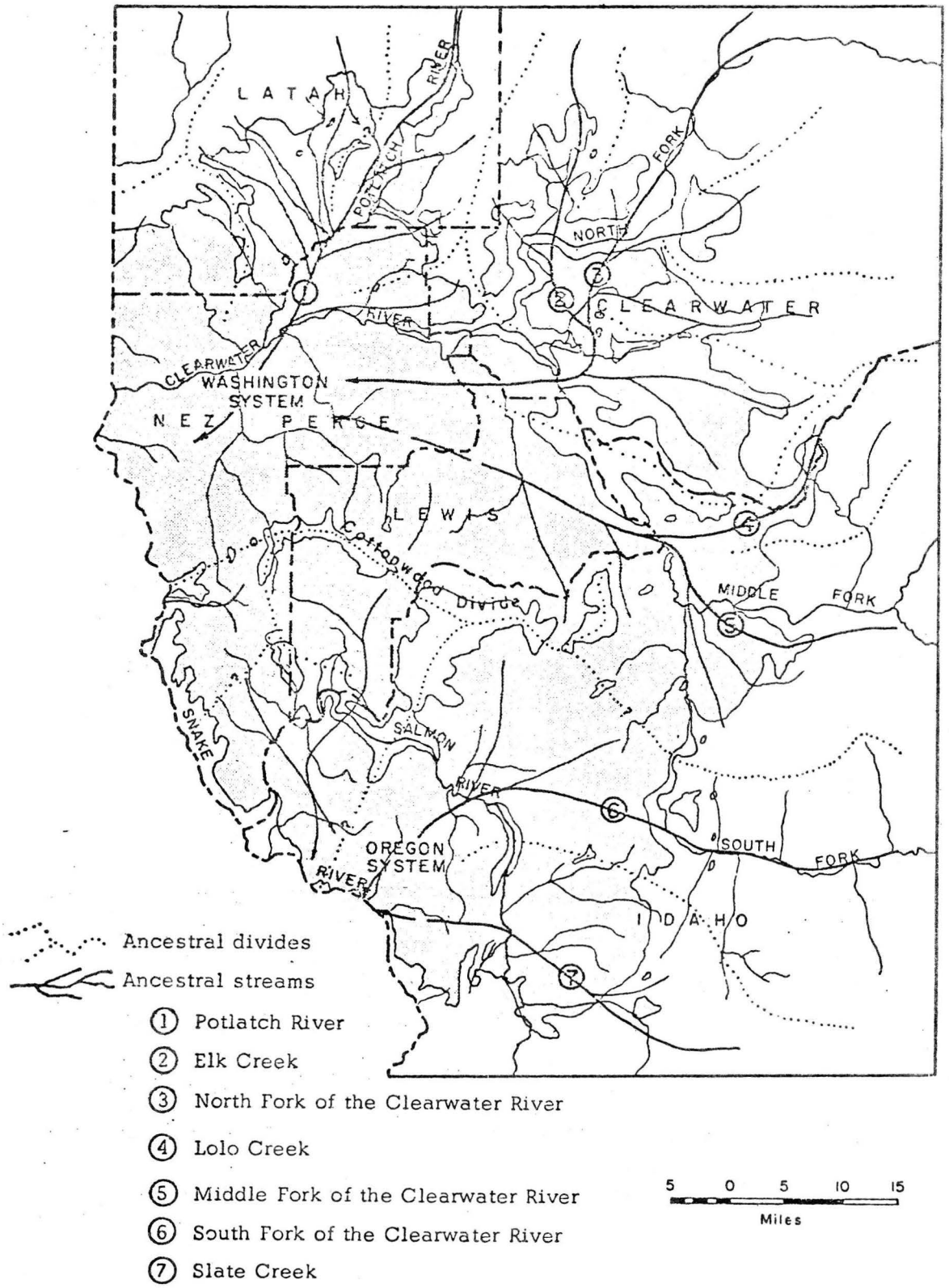


Figure 26. Major structural features in the Pacific northwest.

> depth, and finer texture of the darker colored Gwin soil which produce much more vegetation on slopes underlain by "Upper" Basalt. A difference is also recognizable in the type of natural vegetation on the gentler slopes of the canyons; large pine trees are more common on the Gwin soil." (Bond, 1963)

Within the Columbia River basalt are several interbeds of fluvial and lacustrine sediments. These interbeds are called the Latah formation (not shown on the Grangeville AMS map). The basalts lie on a basement of three units; Belt rocks to the north, the Seven Devils metavolcanics to the south, and the Idaho batholith to the east.

The basalts flowed out onto a surface which had considerable relief and remnants of mountain range peaks are now sticking up through the basalts. These remnants are called Steptoes. The effects of the basalt flows on the ancestral drainage of the Clearwater Embayment are shown in Figure 26 (Bond, 1963). The basalts are gently folded into large anticlines and synclines. Some of the major faults in the basalts are shown on the Grangeville AMS map.

A few small patches of Early Tertiary (pre-Columbia River basalt, Toel) volcanics outcrop at Kamiah Butte and Potato Hill on the Grangeville map. These may correlate with Challis volcanics in central Idaho. There are also a few post Columbia River basalt flows which are found unconformably being on Upper Columbia River flows.

Subsection 18 - Lower South Fork Clearwater Strongly Dissected Lands Subsection

Subsection 18 has been mapped by Kuhns (1974). The subsection is drained by the northwards flowing Johns Creek, a tributary of the South Fork of the Clearwater River. Kuhns recognized two topographic provinces, a relatively flat uplands section and a young steep sided canyons section which he correlated with two geologic sections. The uplands section was the partially dissected "Old Erosion Surface" which was described as: (1) large areas of incompetent, deeply decayed, and weathered rock, (2) large areas of thick soils, (3) the dominance of chemical weathering over mechanical weathering, (4) the rarity of rock outcrops, (5) erosional remnants of Tertiary sediments and (6) rolling topography. The young, steep-sided canyons section, like the South Fork of the Clearwater and Johns Creek are characterized by; (1) abundant rockfall, (2) large areas of fresh outcrop, (3) large areas of incohesive rock structures, (4) oversteepened and unstable slopes, (5) a few slides and, (6) both mechanical and chemical weathering.

Major rock types in the area consist of Belt metasediments (Biotite gneisses, schists and quartzites) and Idaho batholith rocks (quartz monzonite to granodiorite). Kuhns' described a mixed unit of batholith and partially melted metasediments which is common to the entire meta-Belt region of Idaho.

The complex structure in the Subsection trends north-south. Drainages and Basins are fault controlled. A more complete picture of the importance of these north-south faults is given by Reid (1959) (see Subsection 18). Kuhns' mentioned the four fold sets which are described by Reid, Morrison and Greenwood (1973) (see introduction to Section III).

Kuhns' report gives many particulars of interest to geological engineering problems including aggregate description, slope stability and cohesiveness, and mass wasting processes. A section on mineral impact within the area is also included.

Subsection 19 - Elk City Basin Upland Subsection

Subsection 19 centers on Elk City, Idaho. The area is drained by the American River, Newsome Creek, the Red River System, and the Crooked River, all of which feed the South Fork of the Clearwater River to the west. Reid (1959) described the area as being a high, dissected upland with broadly concordant ridge levels. The various drainage valleys are controlled by north-south trending faults. The Tertiary history of the subsection is similar to several of the surrounding subsections and is tied to the "Old Erosion Surface." Reid summarized the geomorphic sequence of Tertiary events as follows.

Postulated Geomorphic Sequence

> "Although no evidence has been found relating to the age of the more recent geologic events in the Elk City region, a geomorphic sequence differing somewhat from those suggested earlier, can be proposed on the basis of present evidence.
>
> The earliest geomorphic event recorded was the formation of an erosion surface of low to moderate relief, with thorough rock decay to considerable depth. Next was an episode of faulting that probably occurred rather rapidly, perhaps with local uplift present only in the vicinity of the Reed Mountain-Pilot Knob-Baldy Mountain ridge, which would account for the abundance of white quartzite boulders present everywhere in the gravels. Relatively rapid uplift seems necessary to account for sufficient local relief to produce the poorly sorted, mudflow-like deposits with their large boulders. From this ridge and others, a blanket of gravel buried the countryside to the east and southeast for some distance beyond Anderson Butte, although judging from the peat beds, the center of the basin may have been near Elk City.
>
> Faulting, following gravel deposition, formed the Newsome Creek and Elk Creek topographic lows and the Elk Summit

and Anderson Butte ridges. One fault, presumably the one along which Newsome Creek now lies, was thrown down to the west so that the gravels on the west side of Newsome Creek are about 300 feet lower than those on the east side. Similar faults under Little Elk Creek and American River are responsible for the uplift of gravels on Erickson Ridge above those west of Little Elk Creek, and for the uplift of the gravels in the head of Baboon Gulch above those on Erickson Ridge. Presumably, further movement occurred along the original faults that formed the basins, for example, along the fault west of Newsome Creek.

Newly formed streams (Newsome Creek, Elk Creek, American River, Red River, and Crooked River in part) gained their courses either as consequent streams in fault-line depressions or as subsequent streams eroding headward along faults; perhaps both mechanisms were locally operative. Evocation of one or the other of these mechanisms seems necessary to explain why each of the major streams lies in a fault zone. Certainly all the streams, if flowing randomly, whether on a continuous gravel surface over the whole region or in individual basins, would not cut down through the gravel in such a way as to flow along a pre-gravel fault zone. This is not to say that the fault zones along which the streams lie are entirely post-gravel; rather it is thought that in post-gravel time renewed movement occurred along some or all of the older faults, some of which have been mineralized in varying degree and are therefore older than the time of ore deposition in this region. Those parts of the streams that lie in strong fault zones where the bedrock has been "softened," have been able to cut rather wide flood plains. The broad meadows of Elk Creek may be broad because the stream is cutting in soft, easily eroded Tertiary gravels.

On the other hand, the lower parts of Red River, American River, and the middle part of Crooked River seem to have gained their present courses through superposition onto the bedrock from the old gravel surface. If superposition is the correct explanation for the course of the middle part of Crooked River across the ridge west of the Crooked River topographic low, the gravels there may have been considerably thicker than 600 feet. Yet, the presence of fault gouge under the creek, even in the narrow part of the canyon, argues against superposition.

The suggestion of a blanket of gravels is not new (Shenon and Reed, 1934a, p. 23), but it is not clear if those authors thought that the gravels were as widespread as here proposed. If gravel has been spread over most of the Elk City region, it is not easy to understand how the

> South Fork gained its present course. Perhaps the syncline formed by the base of the Tertiary gravels is post-gravel and pre-faulting, or contemporaneous with faulting; and perhaps the South Fork is essentially consequent along the axis of the syncline. Though I propose to disregard for the present the problem of how the river got across the postulated Pilot Knob quartzite ridge, Capps' (1941, p. 1) proposal of antecedence, coupled with a suggestion of temporary damming or defeat, suggests a solution.
>
> If one indulges in a bit of pure speculation, it seems possible that the entire north-central Idaho region, extending south of the Salmon River, may have had spread over it a more or less continuous blanket of gravel at some time in or about the Miocene. If so, the present gravel deposits are infaulted remnants of the original gravel blanket. Serious problems arise from this hypothesis regarding the origin of the courses of the major streams of the region.
>
> After the post-gravel faulting, the streams in the Elk City region appear to have developed mature profiles and rather broad flood plains, as in upper Newsome Creek, upper Crooked River, upper Red River, and upper American River. Then, upon a relative lowering of base level, the South Fork cut down rapidly and caused corresponding downcutting and steepening of profile in the lower part of its tributaries, a process still actively under way." (Reid, 1959)

The bedrock in the Elk City area was described by Reid who recognized four major units: (1) Quartzite which may be biotite quartzite, muscovite quartzite (relatively rare), pure while quartzite or feldspathic-biotite quartzite which grades to biotite gneiss. (2) Biotite gneiss which is the most abundant rock in the area. (3) Augen gneiss which is in the southeastern part of the subsection, and (4) granitic rocks which mixed with metamorphic rocks make up more than half the total rock volume. The biotite schists and quartzites are now tentatively assumed to be equivalent to the Prichard Formation (mpC11) and are so designated on the Elk City AMS map (Reid, Greenwood, and Morrison, 1973).

Subsection 20 - Selway Moderately Dissected Land Subsection

Subsection 20 is drained by the Selway River system. Anderson (1956) described the rock types in the subsection as gneisses and schists during a very general reconnaissance of the region. Greenwood and Morrison (1973) show that the Belt metamorphic equivalents mpC11 (Prichard Formation), mpC12 (Ravelli group) and mpC13 (Wallace Formation) continue through this subsection from the Middle Fork of the Clearwater. More detailed descriptions of these units are given in Subsections 19 and 21.

Subsection 21 - Copper Butte Dissected Glaciated Land Subsection

Subsection 21 is drained by tributaries to the Selway River. According to Greenwood and Morrison (1973), rock units within the area include garnetiferous-biotite quartzite, pelitic schists and gneisses, and some minor calc- silicate layers. They tentatively assigned these units to the Ravelli group (mpC12). A band of Wallace equivalent metamorphic rocks (mpC13) extends from northwest to southeast through the central portion of 21. These are described as quartz-biotite-plagioclase gneisses and schists, and biotite-plagioclase quartzites with no calc silicates. The central area of 21 is composed of quartz monzonite of the batholith while the northeastern third is a Tertiary pluton (Tei3). The Tertiary pluton is similar in description to the Sawtooth batholith (Section I, Subsection 12).

Major linears show the standard pattern of north-south trending faults which control drainages and valley development. The western contact of the Tertiary intrusive may also be fault controlled as it lies on a north-south linear.

Subsection 22 - Running Creek Streamcut Land Subsection

Subsection 22 contains the drainage of Burning Creek where it flows into the Selway River and the northwest flowing segment of the Selway River itself. Small outcrops of Wallace equivalent metasediments (mpC13) occur throughout the area but most of 22 is underlain by quartz monzonite (Ki3C) of the batholith or the northeastern part of a Tertiary pluton (Tei3).

A major linear which is the extension of the Bargamin Creek fault (Subsection 12) goes through 22 along with several smaller parallel linears. The main linear can be traced from the Salmon River north through about one third of the Hamilton AMS map. This long linear (and parallel ones) conincide with offsets in the Salmon River, the Selway River, and the East Fork of Moose Creek. The north-south flow direction of the Selway River appears to be a possible north-south offset of the older westward flow direction. Greenwood and Morrison (1973) discussed the general structure of the Selway-Bitterroot Area and mentioned the possibility of transcurrent movement on the long faults and the significance of these movements. Their discussion follows.

Structure of the Selway-Bitterroot Wilderness Area

General

> "The structures observed in the Selway-Bitterroot Wilderness Area are generally compatible with the structural history of the Clearwater orogenic belt as a whole. The structural features which have been observed in the metamorphic rocks of the wilderness area will be described in order to show this similarity. The structure of the igneous rocks will also be described briefly."

Major Folds

"A flexure of regional extent in the northwest trending regional strike of foliation occurs near Green Mountain.

This sigmoidal feature has been interpreted as being the result of right-lateral strike-slip motion on a northwest trending zone in the basement (Greenwood and Reid, 1969). The sigmoidal structure bends lineation related to second generation folds, and is, in turn, sheared by cataclastic folds, and intruded by quartz monzonite. Therefore, the sigmoidal folding could be from latest Precambrian to Jurassic in age."

Minor Folds

"Isoclinal folds of microscopic to outcrop size have been observed in the wilderness area. These folds are of three general kinds: (1) knife-sharp isoclinal folds of compositional layering (S_o) with the production of an axial-plane schistosity (Si) throughout; (2) tight isoclinal folds of schistosity (Si) with the production of an incipient axial-plane schistosity (S_2) in the noses of folds; and (3) moderately tight isoclinal folds of schistosity without production of a new schistosity parallel the axial-plane of the fold. Fold types one and two have been observed in the same outcrop and commonly diverge in axial trend by 20-30°. Folds of type 3 have been observed to refold type 2 folds in outcrops on the Salmon River near Whitewater Ranch. North-south trending cataclastic folds have been observed along the Lochsa River and on the trail north of Sheep Hill. These structures differ in style and orientation and appear to represent four distinct periods of deformation."

Faults

"The fault pattern of the Wilderness Area is very complex. Deep weathering in fault zones and the lack of marker beds makes directions and magnitude of movement difficult to determine. However, certain major trends can be discerned by the length and degree of development of certain linear features. One of the most prominent linear features is that of the N-S trending Bitterroot front just east of the border of Montana and Idaho. Within the Idaho portion of the wilderness area, just west of the front, there are numerous thick, brecciated and iron-stained zones which trend north-south. The simplest explanation is that these zones were produced by dip-slip motion during the uplift of the Bitterroot front. However, the degree of brecciation

and alteration, the straightness, and length of the zones suggest the possibility of transcurrent movement. North-trending strike-slip faults are reported in the Salmon, Idaho, area to the south of the wilderness area (Ruppel, 1967).

West of the N-S trending zones the most prominent linear zones are aligned in a northeast direction. One such zone follows the valleys of Bargamin Creek and Running Creek for a distance of over 15 miles. In this zone the rocks are highly sheared and intruded by many dikes. These dikes are, in turn, sheared suggesting recurrent motion in the shear zone. The northeast linear features parallel the trend of strike-slip fault in the Atlanta area of central Idaho (Anderson, 1948). In the Atlanta area, Anderson described northwest trending gash veins and faults as tension features related to northeast, right-lateral transcurrent motion." (Greenwood and Morrison, 1973)

THE GEOLOGY OF SECTION IV

LOCHSA UPLANDS SECTION

Section IV is subdivided into six subsections. Most of the section lies in the drainage of the Selway and Lochsa Rivers which join the Middle Fork of the Clearwater River to the west. The eastern two thirds of Section IV is underlain by rocks of the Idaho batholith whereas the western third is composed of batholith and Precambrian metasediments. Subsection 2 extends to the northwest of the Lochsa River and includes an extremely complex system of Precambrian metasediments, batholith rocks and a few of the eastern outliers of Columbia River basalt.

The Lochsa uplands subsection has a geomorphic history similar to the Elk City area. The development of a mature topography in the early Tertiary was followed by regional uplift, faulting, and dissection (see Reid's discussion, Section III, Subsection 19).

While similar in topography and geomorphic history, the Lochsa section has a markedly different tectonic pattern as compared to sections to the north and south. The linear pattern which trends north-south or northeast (south of Subsection IV) changes to a strong northwest orientation. This change in tectonic pattern is associated with the Trans-Idaho Discontinuity. The discontinuity is related to major tectonic provinces to the north associated with the Lewis and Clark thrust zone. Norman Day (1974) has formed several new ideas about the discontinuity which follow verbatim (except for editorial remarks for clarity).

A New Look at the Trans-Idaho Discontinuity

> "This discussion will deal primarily with three topics: first, some of the significant transverse structures of the Pacific northwest; second, structural aspects of air-photo linears in Idaho and their relationship to bedrock and aeromagnetic maps; third, comments concerning sense of movement, age, and width of the Trans-Idaho Discontinuity.
>
> The Pacific Northwestern States, particularly Idaho, are cut by several northwest trending strike-slip fault zones (Figure 26a). The Lewis and Clark fault zone, located in northern Idaho and western Montana, is comprised of three main fault zones: the Hope, Osburn, and St. Joe fault zones. Recent work by Cady (1974), in the Coeur d'Alene Mining District, indicates a sequence of left lateral, right lateral, normal and right lateral faulting within the Osburn fault zone. With equally detailed studies along the Hope and St. Joe fault zones similar complex histories of movement may be demonstrated. The Olympic-Wallowa lineament has been described as showing either left or right lateral displacement by different workers; although there is not complete agreement on whether the lineament represents a fault.

In 1968 Yates suggested that a major northwest trending fault or zone of faults with left-lateral displacement extends across the Pacific northwest between the Lewis and Clark structural zone and the Olympic-Wallowa Lineament and proposed the name Trans-Idaho Discontinuity for the feature. Yates based the location and sense of movement on a series of apparent lithologic offsets. These offsets occur in:

1. The eastern limit of Mesozoic Alpine ultramafic rocks.

2. The eastern limit of Permian and Triassic volcanic rocks.

3. The isopachs of Middle and Upper Cambrian miogeosynclinal sedimentary rocks.

4. A belt of Upper Precambrian volcanic rocks.

The Lewis and Clark zone and the Olympic-Wallowa lineament both fail as possible candidates to account for these offsets. However, a northwest trending fault or zone of faults with left-lateral displacement, as proposed by Yates, provides a viable explanation.

In 1972, a reconnaissance structural study of 8,500 square miles in the rugged mountains of north-central Idaho (Figure 27) using airphoto linears was undertaken. For this study linears from three types of imagery were used. These were conventional black and white aerial photography at a scale of 1:62,000; two channel thermal infared (3.0-4.0 and 8.5-11.0) at a scale of 1:200,000; and erts imagery, all four bands, all reproduced at a scale of 1:500,000. After interpretation of each separate type of imagery for linears, a composite map of linears from all three types of imageries was compiled (Figure 28). The heavy lines are linears which match up on at least two of the imageries.

As can be seen from Figure 28, there are several distinct linear trends and linear patterns. First, is a zone of northwest trending linears which terminates against a wide zone of very long north-northeast trending linears; second, is a narrow and less distinct zone of northwest trending linears; third, is a wide zone of curvilinears which appears to represent drag features associated with left-lateral movement along the first northwesterly zone of linears; fourth, is an area

where the pattern of linears (this includes density of linears, length of linears, number of linear intersections, and number of directions of linears) is different than in any other region within the study area. Another unique feature of the region is that it is highly mineralized, while only scattered mines are found within the rest of the study area.

Comparison of airphoto linear trends with aeromagnetic data demonstrates a positive correlation between the two (Figure 29). Perhaps the most striking feature of this aeromagnetic map is the northwest trending ridge of anomalously high magnetic readings. This ridge corresponds exactly with the major zone of northwest trending airphoto linears.

The aeromagnetic ridge is broken and shows some evidence for left-lateral displacement at the point where the zone of very long north-northeast linears intersect it. Zeitz and others (1971) believe this ridge may be a steeply dipping dikelike body of more mafic composition than the surrounding rocks. Whatever its origin, it appears to be a major structural feature.

The next illustration is a composite geologic map of the study area (ed. note - see Hamilton AMS geologic and linear map). Comparison of the composite geologic map with the composite airphoto linears map shows that many airphoto linears correlate with mapped bedrock contacts and faults and that the two zones of northwest trending linears correspond to local margins of the Idaho batholith and several northwest trending faults. There is also an indication that two of the tertiary intrusions may be offset left-laterally from one another. A series of tertiary intrusions are seen to be aligned along a northwest trend in the area of the major northwest airphoto linear zone; there is also a minor elongation of some of the plutons in a northwesterly direction. The location of the two large tertiary plutons is exactly in the spot where the aeromagnetic ridge is broken by the zone of very long north-northeasterly trending airphoto linears. You will also notice that these plutons show a strong elongation in a north-northeasterly direction; apparently, emplacement of the plutons was controlled by north-northeasterly fractures. The association of numerous geothermal springs with these north-northeasterly linears is also suggestive of localization of relatively recent igneous activity along the north-northeasterly fractures.

The foregoing data suggests that this portion of Idaho has undergone major structural disturbances, perhaps subcontinental in proportion. The Trans-Idaho Discontinuity seems to have played a significant role in these disturbances.

While this may be only a photogeologic study, several comments concerning the nature of the Trans-Idaho Discontinuity can be drawn from the data. First is width. The Discontinuity appears somewhat wider than Yates first proposed. This author favors placing the northern limit of the zone at the northern boundary of the Idaho batholith. Following Yates, the Lewis and Clark zone is excluded from the Discontinuity because it cannot account for the offsets which suggested the presence of the Discontinuity. The southern limit, although out of this study area, is believed to be associated with the Florence-stibnite mineral belt of Green's 1972 (Figure 30). This mineral belt is a northwest trending band of mineral sites and linear features.

Second is apparent age. The correlation of airphoto linears and the borders of the Idaho batholith in the study area leaves open the possibility that the emplacement of this portion of the Idaho batholith may have been controlled by pre-existing northwesterly structures. Comparison of northwest and north-northeast trends in the Precambrian belt basin, during Ravalli time, with airphoto linear trends (Figure 31) further implies that in origin these linear trends may well represent Precambrian age structures; possibly representing transform faults associated with a late Precambrian plate separation as proposed by Stewart (1972). Elongation of tertiary intrusions suggests that the most recent activity is Cenozoic in age and was mostly along north-northeasterly fractures.

Third is a sense of movement. The zone of airphoto curvilinears and the lithologic evidence compiled by Yates implies left-lateral displacement. However, heat flow studies within the Idaho batholith by Swanberg and Blackwell (1973) indicate right-lateral displacement of a body of muscovite quartz monzonite.

A certain amount of vertical displacement is also possible if Harrison's proposed Precambrian belt island existed in this area. Without significant and detailed field work in this region, and accurate structural history is not possible. This author believes that movement along, or perhaps more properly within, the Discontinuity has been sporadic and of varying senses of movement.

In conclusion, I would like to summarize my impression of the Trans-Idaho Discontinuity.

1. It is roughly a northwest trending zone of predominantly strike slip faults with varying direction of movement which, over time, has resulted in a net left-lateral displacement.

2. Some of the faults are of Precambrian age and have had sporadic and recurring activity into the Tertiary.

3. In Idaho, the zone is approximately 100 miles wide and gradually narrows in a westerly direction where it is covered by Columbia River basalts. Therefore, the nature of its western extension cannot be verified.

4. In western Idaho and eastern Washington where the Discontinuity is covered by the Columbia River basalts, Taubeneck (1973) has measured trends within Pre-Tertiary rocks and has found the Discontinuity to strike north 70° west.

5. Lastly, the correlation of airphoto linears with aeromagnetic anomalies suggest that basement trends have controlled or at least influenced intrusion, the Tectonic history of younger rocks and possibly sedimentation, within this portion of Idaho." (Day, 1974)

Subsection 1 - Selway-Lochsa Canyon Land Subsection

Subsection 1 includes most of the main drainage of the Selway and Lochsa Rivers which, like most of the rivers in northern Idaho, are deeply entrenched. Much of the western part of Subsection 1 is covered by a series of Precambrian Belt metasediments including rocks equivalent to the Prichard Formation (mpC11), the Ravalli Group (mpC12) and the Wallace Formation (mpC13). The Prichard Formation is composed of quartzites and schists, the Ravalli Group, quartzites, various schists, and some calc silicates and the Wallace Formation schists and gneisses with frequent calc silicate units (Wallace is the most calcareous of the Belt Supergroup). These rocks are aligned in a northwesterly trend which changes to a curvilinear pattern in Subsection 2. An interesting feature of Subsection 1 is the offset in the Lochsa River which may show left lateral strike slip movement on a north trending fault system.

To the northeast along the Lochsa River are several outcrops of the Idaho batholith (coarse quartz monzonite, Ki3c) while southeast along the Selway the border of Subsection 1 intersects a Tertiary intrusive which was discussed in the introduction.

Subsection 2 - Clearwater Uplands (Old Erosion Surface) Subsection

Subsection 2 extends from the western portion of the Lochsa River, northwest to the North Fork of the Clearwater and slightly beyond. The area is conveniently discussed in two parts; the south and northeastern portions from the Lochsa River to the Clearwater/Idaho County line and the northwestern part from the county line to the North Fork of the Clearwater.

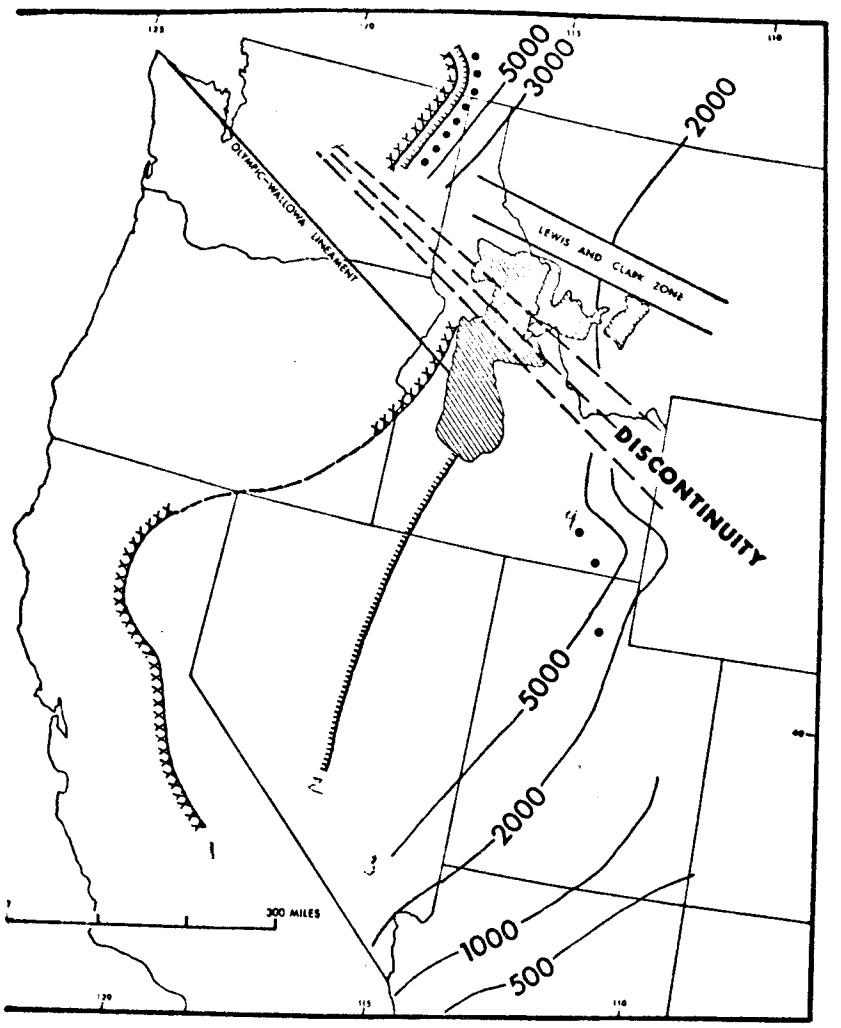


Figure 26a. Major structural features in the Pacific northwest

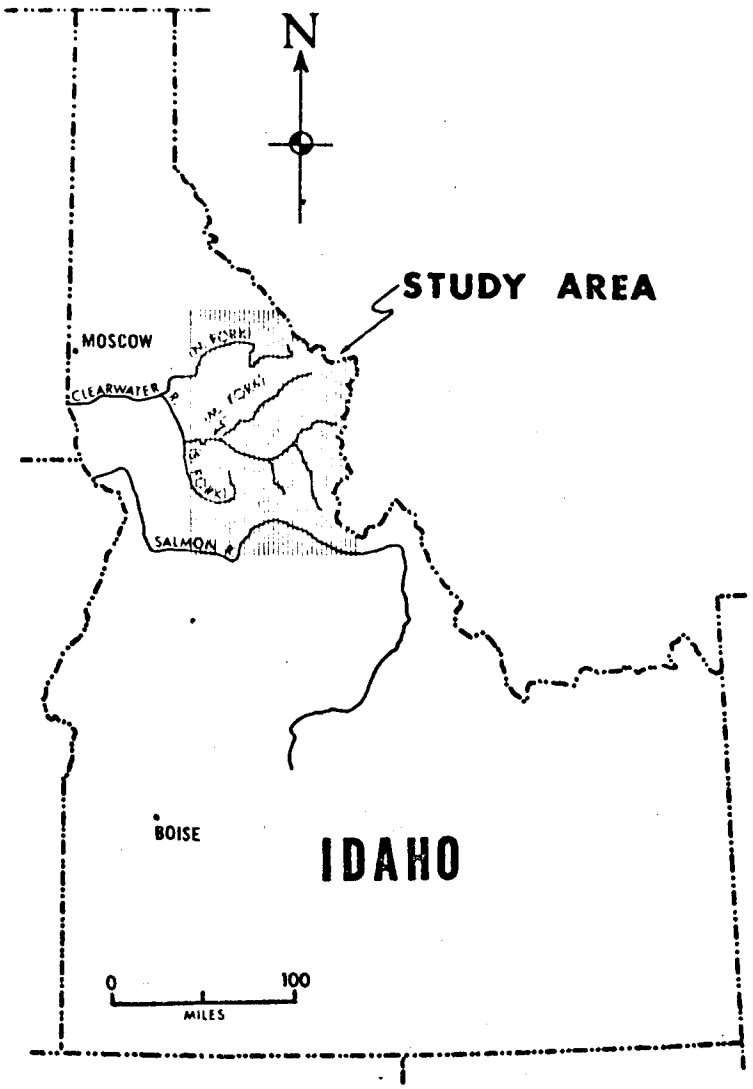


Figure 27. Location of air photo study area (Day, 1974)

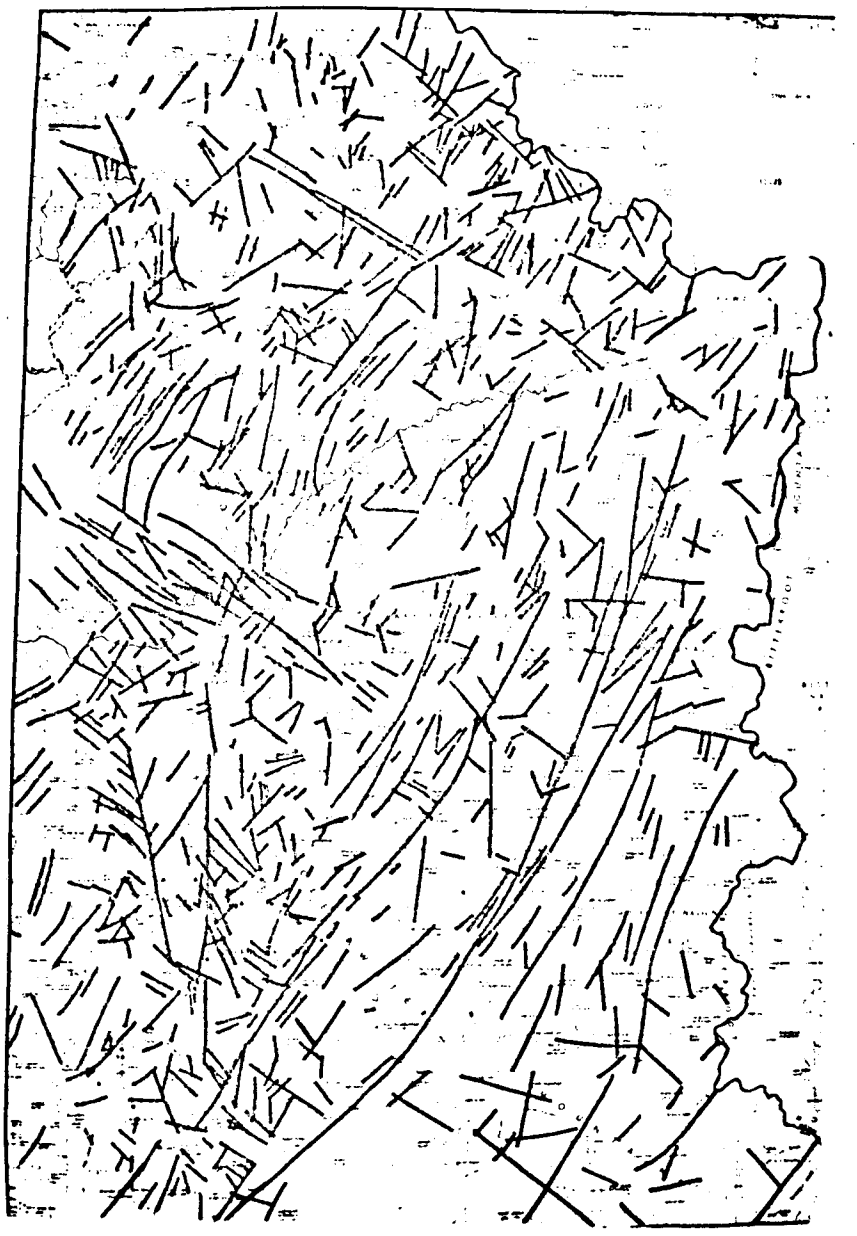


Figure 28. Composite linear map based on Conventional Infrared, and ERTS photography (Day, 1974)

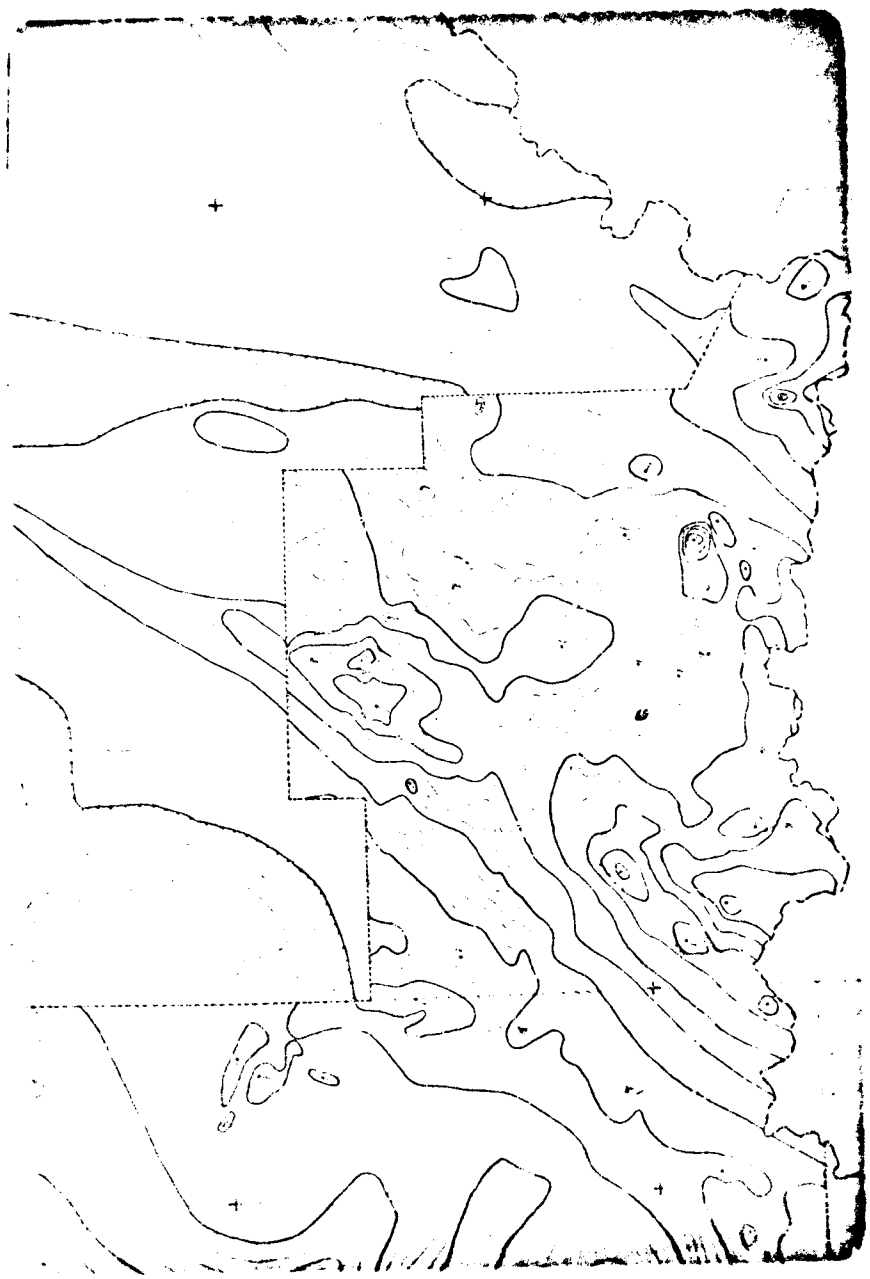


Figure 29. Aeromagnetic map of the area shown in Figure 28 (Zeitz, et. al., 1971)

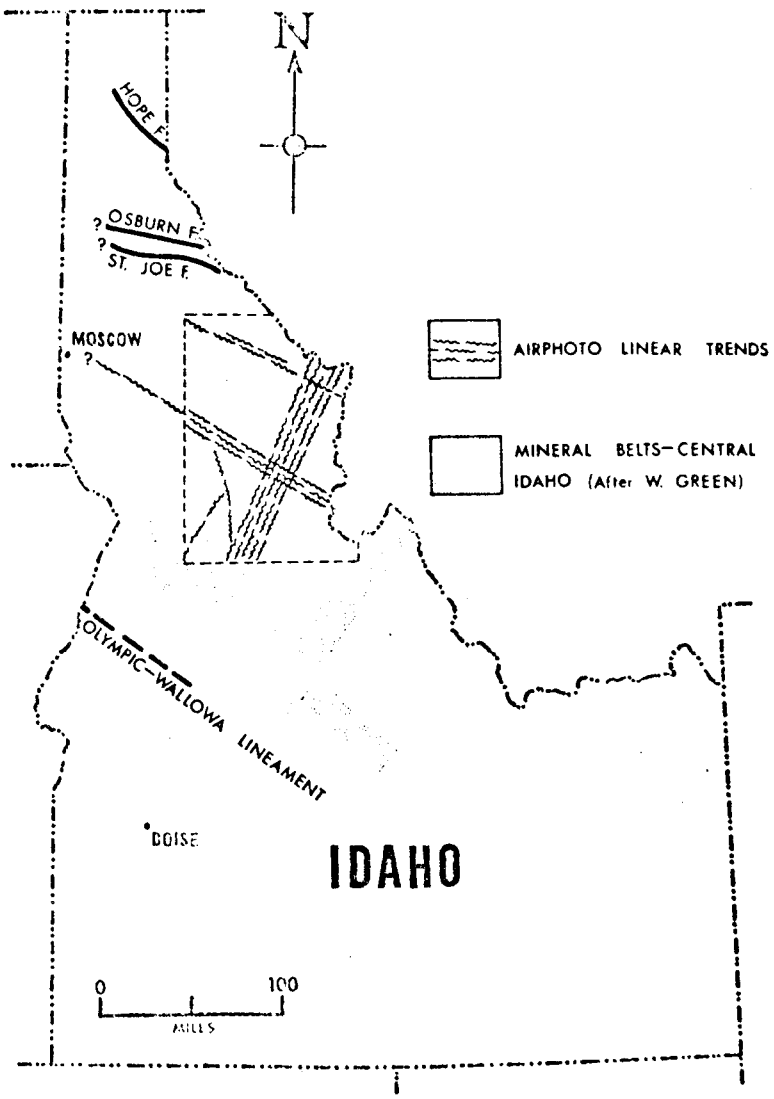
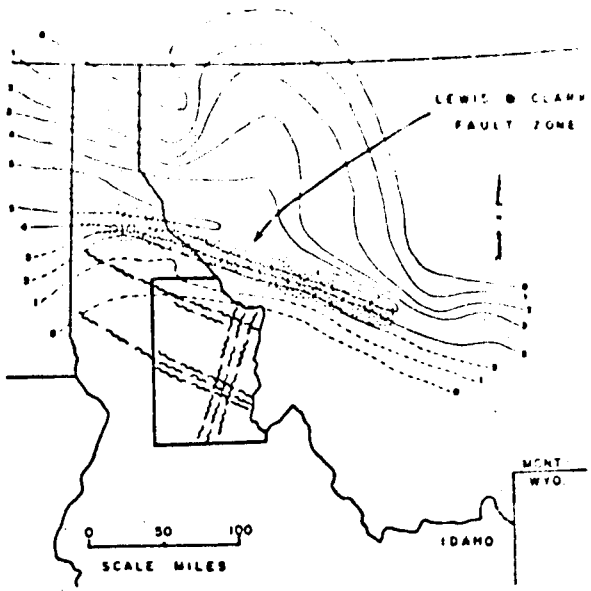


Figure 30. Composite map showing linear and Mineral Belt trends (Day, 1974)



(Modified after R. Boyce)

Figure 7 - Isopach map of Combined Revett-St. Regis (data from Harrison, 1972). Contour interval is 1000 Feet.

Figure 31. Composite map showing Precambrian Belt Basin (Ravalli time) and linear trend (1974)

The drainage pattern in Subsection 2 is different from other drainage patterns in the section. The southwestern part of 2 drains into Lolo Creek instead of into the Middle Fork of the Clearwater. Drainages flow almost due south to feed Lolo Creek whereas the overall pattern of the major drainages in the section is westward flow. This subsection is characterized by curvilinears which may be faults controlling the north-south drainage pattern (see Hamilton Linear maps).

Major lithologies in the south and northeastern parts of the subsection are the Idaho batholith (almost entirely quartz monzonite, Ki3c or Ki3ch) with a few areas of undifferentiated Precambrain metasediments (mpCl) along the southern border.

The northwest part of the Subsection is drained by Orofino Creek and the north Fork of the Clearwater River. This area has been mapped in great detail by Hietanen (1962). Figure 32 is a generalized geologic map of this area showing structural symbols. Note that the general trend of fold axis and faults is to the northwest. The curvilinear trend of Subsection 2 (southern part) has turned to a north west trend. Figure 32 also shows the index minerals of the various metamorphic facies in the Belt rocks. As the batholith is approached the grade of metamorphism increases with index minerals Kyanite-Andalusite(lower temperatures) grading to sillimanite (high temperature). The structures in this region, like Belt structure to the south, are multiple deformed which makes interpretation difficult. The style of deformation is similar to the discussion of Reid, Greenwood, and Morrison (1973, see general discussion of Section III). The very northern part of the Subsection contains an anorthosite body. Anorthosite is a rock composed essentially of only calcic plagioclase. Later descriptions of the Bitterroot Mountains (Section V) include anorthosite bodies as well as most of the other features of Subsection 3 (following).

Subsection 3 - Craigs Alpine Glaciated Lands Subsection

Subsection 3 is a large area of Idaho batholith (quartz monzonite, Ki3c). It is drained by the Lochsa River to the north and the Selway River to the south. Almost all of the highest elevations within the subsection are around 8,000 feet. These mountain tops and ridges represent the "Old Erosion Surface" which has been frequently mentioned. Elevations in the Lochsa River bottom are around 3,500-4,000 feet. Subtracting maximum and minimum elevations shows that about 4,000 feet of uplift has occurred since the Tertiary surface formed.

Subsection 4 - Moose Creek Headlands Subsection

Subsection 4 is drained by Moose Creek which flows from east to west but which has a couple of north-south steps or offsets. One of these offsets coincides with the "step" in the northern contact of the Tertiary intrusive in the southern part of 4. Another offset conincides with a large offset in the Lochsa River to the north.

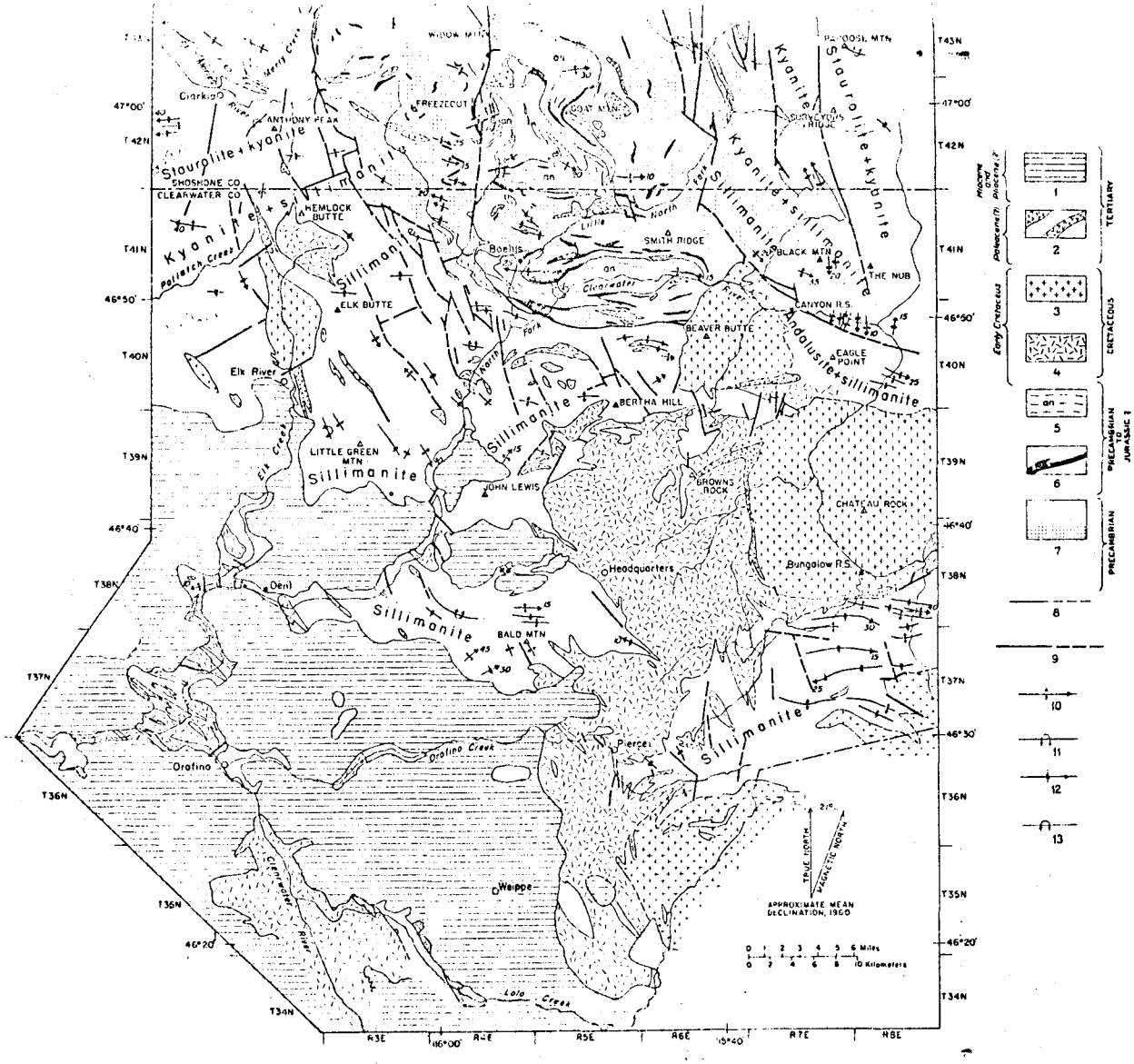


Figure 32. Sketch map showing the distribution of major rock types and some of the index minerals of the metamorphic facies in western Clearwater County, Idaho, with legend (after Hietanen, 1961).

The linear map of this section shows a north-northeast trend for Subsection 4. The drainage offsets may be due to strike slip movements along these linears. With the exception of the Tertiary pluton, the rest of Subsection 4 is quartz monzonite of the batholith as mapped by Greenwood and Morrison (1973).

Subsection 5 - Bitterroot Glacial Modified Subsection

Subsection 5 is drained by Moose Creek and its tributaries in the south and by tributaries of the Lochsa to the north. Most of the area was mapped by Greenwood and Morrison (1973) as Idaho batholith with a few stringers of Precambrian metasediments in the southern and central areas. Parts of the subsection are mapped as foliated quartz monzonite which is probably indicative of the nearness of the eastern batholith border. The presence of several small quartz diorite stocks was also noted.

Much of the northern part of 5 was mapped by Nold (1968). His discussion should be compared with Reid, Morrison, and Greenwood (1973, Section III general discussion) as it is quite similar. Nold summarized the geology of the north part of 5 as follows (Table 3).

> "The area encompasses the northeastern border zone of the Idaho Batholith, its associated plutons, and the high- to low-grade regional metamorphic rocks into which they were emplaced. The emtamorphic rocks are equivalents of the Prichard Formation, Ravalli Group, and the Wallace Formation, all from the lower portion of the Precambrian Belt Supergroup. A geologic history of the area appears as Table 3.
>
> Metamorphism in the area appears to be mostly regional, with contact metamorphism visible around outlying stocks and batholiths where they come in contact with lower grade regional metamorphic rocks. The grade of regional metamorphism ranges from greenschist facies (biotite zone) in the northern part of the map area, to the upper amphibolite facies (sillimanite zone) in the southeast, adjacent to part of the Idaho batholith. The principal metamorphic rock types present in the area are pelitic schist, quartzite-quartzitic gneiss, and calc-sillicate gneiss, respectively correlated with the Prichard Formation, Ravalli Group, and Wallace Formation.
>
> The structure of the area is complex. The region has undergone at least four deformations. The earliest is an isoclinal folding with axial plane schistosity, probably concurrent with the main prograde regional metamorphism in the area. This deformation is defined on the basis of mesoscopic folds and no macroscopic folds definitely of this type were noted. These earlier folds and their axial plane schistosity are folded by an obvious macroscopic fold-set trending northwest to west. These

folds appear to be later than a southwest-trending macroscopic fold-set which is most obvious in the southeastern part of the area. In addition to the three deformations outlined above, the area has probably undergone at least one phase of steep open folding (Brian White, personal communication, 1968).

The Bitterroot Range and areas to the west appear to have been uplifted somewhat as a block along the east-west-trending Lolo fault on the north and the north-south-trending Bitterroot faults on the east. The most prominent fault-trend within the map area is northeast to north. These faults are best defined where they displace the contact or the Lolo Hot Springs batholith.

The largest igneous bodies in the area are the Skookum Butte stock, the Lolo Hot Springs batholith, and part of the Idaho batholith, the main body of which lies to the southwest. The Skookum Butte stock ranges in composition from diorite to quartz monzonite, being composed mainly of quartz diorite and granodiorite. Contact metamorphic rocks associated with this body, range in grade up to the hornblende hornfels facies. The Lolo Hot Springs batholith is composed of granite and quartz monzonite, and appears to be a shallow, epizonal intrusive. It is intruded into low-grade regional metamorphic rocks and probably represents the youngest igneous pluton within the area. The Idaho batholith in this area is composed mainly of granodiorite and quartz monzonite. It appears to be more deep-seated than the plutons discussed above. Though the contact is sharp on the scale of a specimen or an outcrop, the contact zone is quite wide (about 15 miles wide along U. S. Highway 12 west of the map area) and consists of a series of sheets of metamorphic rocks interlayered with sills of igneous rocks. It is intruded into regional metamorphic rocks of amphibolite facies which show no obvious effects of contact metamorphism.

The Idaho batholith, in this area, is spatially associated with high-grade regional metamorphic rocks. At least three interpretations of this association are possible. They are:

1. The batholith caused the regional metamorphism; that is, the regional metamorphism is a large scale contact metamorphism.

2. The regional metamorphism developed the magma for the batholith by anatexis.

3. The regional metamorphism is earlier (Precambrian?) than batholithic intrusion (Mesozoic?) within the area, and unrelated to it.

Geologic History as Evidenced in the Area

Era	Period	Sedimentation-Metamorphism-Intrusion	Deformation
Cenozoic	Quaternary	glaciation, erosion, and depositon to develop the present topography	high-angle northeast- to north-south-trending faults
Cenozoic	Tertiary	erosion and deposition accompanying faulting	east-west-trending faults, offset by north-south-trending faults
Cenozoic	Tertiary	Crooked Fork plug Lolo Hot Springs batholith	
Mesozoic	Cretaceous	Igneous Intrusion: Skookum Butte Stock (contact metamorphism up to hornblende hornfels); White Sand Creek stock; Brushy Fork stock; Idaho batholith	? steep-axis open folding bending schistosity, possibly concurrent with igneous intrusion.
Mesozoic	?	minor retrograde regional metamorphism throughout the area, mainly development of chlorite from biotite	northwest-trending similar-style folding; bends schistosity southwest-trending, similar-style folding; bends schistosity
Mesozoic	?	prograde regional metamorphism ranging from greenschist facies (biotite zone) to the amphibolite facies (sillimanite zone)	similar-style (?) folding with axial plane schistosity
Mesozoic	?	intrusion of basic dikes and sills which are later metamorphosed to orthoamphibolites and orthogneisses	
Prec		deposition of the Belt Supergroup rocks: Prichard Fm., Ravalli Gp., and Wallace Fm.	

Table 3 - Geologic history of the northeastern border zone of the Idaho batholith after Nold (1968)

> The first possibility is considered to be unlikely because of the discordance of the batholith to regional metamorphic facies within the area. If the batholith had caused the metamorphism, there should be a closer correspondence of the higher grade rocks to its borders. In addition, the strong schistosity in the gneisses and schists does not suggest development by large scale contact metamorphism.
>
> The third possibility appears unlikely because of the lack of medium- or high-grade regional metamorphism elsewhere in lower Belt Supergroup rocks, except in the area of the batholith.
>
> The remaining relationship, that the batholithic magma was developed by metamorphism, seems most likely considering the field relations within the map area. The main evidence for this relationship is that the Idaho batholith is discordant to the metamorphic facies within the area. After development by anatexis at deeper levels, the batholithic magma apparently cut across the facies during its rise and emplacement.
>
> Although the regional metamorphism is thus inferred to be most likely Mesozoic in age, the evidence is weak and the question is, in fact, unresolved." (Nold, 1968)

Alpine glaciation in the Bitterroots would have provided much of the valley fill material.

Subsection 6 - Lolo Rolling Upland Subsection

Subsection 6 is drained by Lolo Creek and its tributaries which flow east to the Bitterroot River, south of Missoula, Montana. Langton (1935) and Nold (1968) mapped most of the subsection. Precambrian sediments (Ravalli group, pC_{12} and Wallace Formation, pC_{13}) cover most of the area but some metamorphosed Prichard Formation (mpC_{11}) occurs in the southeastern part of the area and Idaho batholith outcrops in the southwestern part.

The structure in this subsection is quite complex and is more conveniently discussed in Section V. Thrust faulting has greatly complicated the geology of this region.

THE GEOLOGY OF SECTION V

BITTERROOT MOUNTAIN SECTION

Subsection 1 - East Slope Bitterroot Range Subsection

Subsection 2 - West Slope Bitterroot Range Subsection

The Bitterroot Mountains are a north-south trending range with peak elevations over 10,000 feet. The range marks the Continental Divide as well as being the boundary for most of its length between Idaho and Montana.

The eastern side of the Bitterroot Mountains contains the Bitterroot Valley which is drained by the Bitterroot River which flows into the Clark Fork near Missoula. The valley, while not part of the included subsections, contains at least 1,800 feet of fluvial, glacial and lacustrine sediments near Corvallis (Langton, 1934).

The Bitterroots may be divided into three main provinces, 1) the Belt Province, 2) the Batholith province, and 3) the Basement province (this term should be used with caution as the southern metamorphics probably are not basement material). The Belt province is in the northern part of the range, the Batholith province is in the central area, and the Basement province lies to the south as can be seen in Figure 33.

The most interesting and controversial part of the range is the remarkably straight eastern border which is over 65 miles long. The slope of this eastern border is very even and is also a subject of controversy. The main question is whether this straight line represents a fault or not and both sides of the argument will be presented shortly after a discussion of rock types and structure in the area, as these aspects are part of the problem.

Examination of the Hamilton AMS map shows that the sequence of rocks from the west to the east in the section is; 1) Idaho batholith granitics (quartz monzonite and granodiorite), 2) the "younger gneiss" of Langton (1935) which equals the zone of weak cataclasis of Chase (1973) and a gneiss unit of Winegar (1974), and 3) a zone of "older gneiss" (Langton), zone of strong cataclasis (Chase) or the "frontal zone gneiss" of Winegar. The northern part of the area which is best typified by Section IV, Subsection 6 and Section VI, Subsection 4 is covered with a thick sequence of relatively unmetamorphosed Belt while the southern part (best example is Section III, Subsection 9) is covered with highly metamorphosed Belt equivalents.

Figure 34 is a simplified map of the major structures in the Bitterroot Range. The major structures in the range are a series of north-south oriented thrust faults which bend westwards near Missoula. The structural trend is then northwest through the Coeur d'Alene region and then again north-south through the panhandle of Idaho.

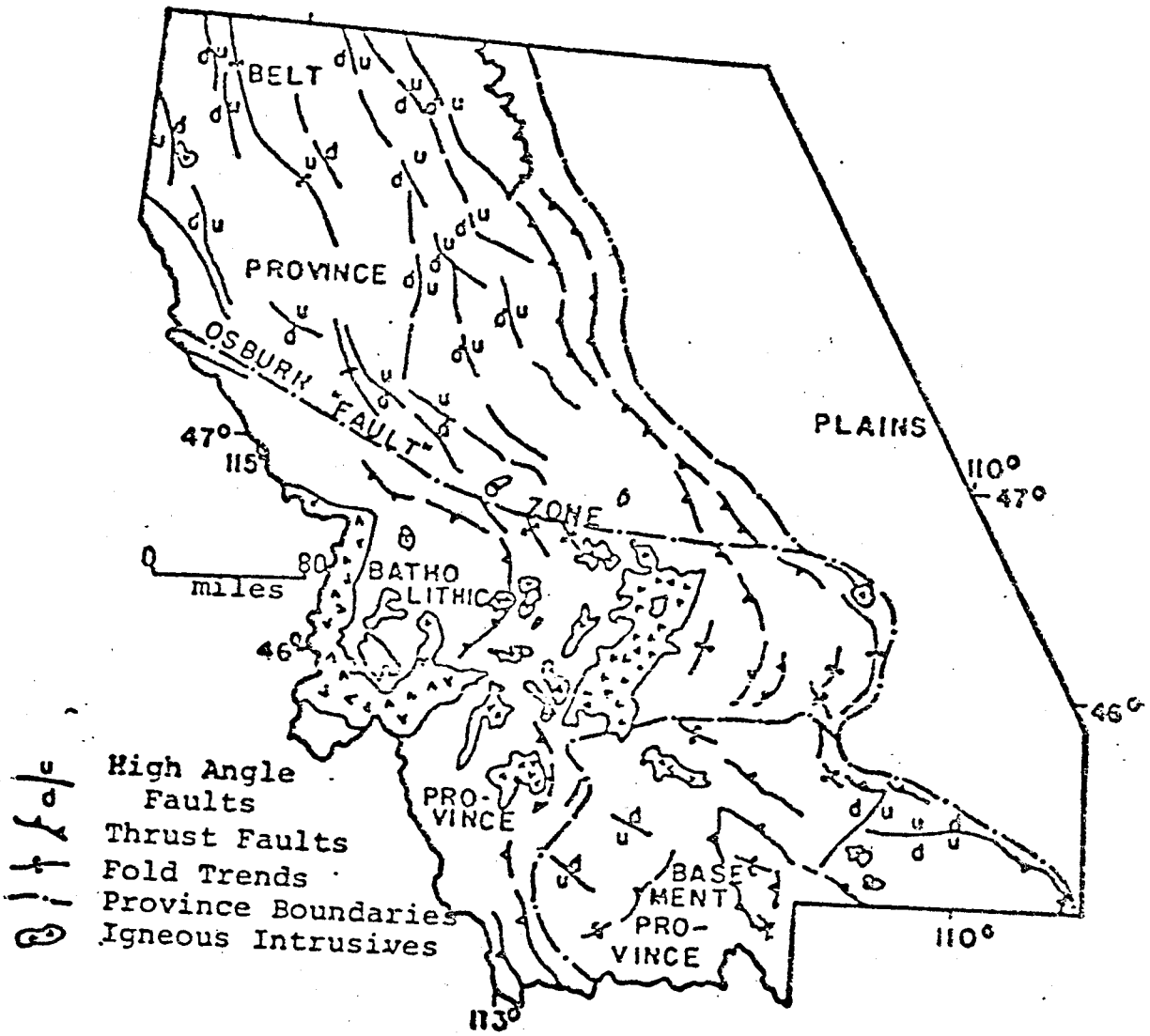


Figure 33. Generalized tectonic map of western Montana showing major provinces (McMannis, 1965)

A series of what were interpreted by Langton (1935) to be truncated spurs (usually indicative of a fault) were observed along the eastern edge of the Bitterroot Range. In addition, normal faults, slickensides, and other evidence of normal faulting were noted. Lindgren (1904) who first postulated the normal front range fault estimated 4,000-6,000 feet of fault throw at a uniform east dipping angle of 15° - 35°. The "older gneiss" which has strong cataclasis was believed to be associated with this fault. Lindgren believed that the scarp producing fault was a normal fault. Langton reinterpreted and added new data and concluded that the Bitterroot fault was a thrust fault that later developed a scarp by normal faulting associated with emplacement of the batholith. Langton's sequence of the structural development of the Bitterroot Valley follows. Figure 35 is a hypothetical cross section which uses Langton's ideas.

> "We may tentatively summarize the structural development of the Bitterroot Valley as follows:
>
> 1. Development of the large north-south trending Bitterroot thrust fault, dipping eastward and probably contemporaneous with the upthrust faults bounding the linear ranges in northwestern Montana.
>
> 2. Batholithic intrusion within the Bitterroot Range, with contemporaneous development of the series of overthrust faults and tear faulting along the older faulted and weakened zone, together with the dragging of the gneiss and bending of the faults to the south.
>
> 3. Normal faulting. This produced, at least in part, the slickensides and a granulation of the gneissic zone by sinking of the valley and uplift of the range along the zone which had been weakened previously by thrust faulting, intrusions, and tearing.
>
> This complex history perhaps in part accounts for the considerable subsidence of the floor in a valley so narrow that its fill, ordinarily, would be easily supported by the strength of the earth's crust."
> (Langton p. 55)

Note that the thrust faults were thought to have been formed by the intrusion of the batholith. Evidence for this is the concentric nature of the thrust faults which seem to wrap around the batholith outcrop area.

Ross (1950) thought that some faulting had occurred along the range front but believed that most of the cataclastic gniesses were formed by granitization of Belt sediments. Ross's discussion and cross section (Figure 36) follow.

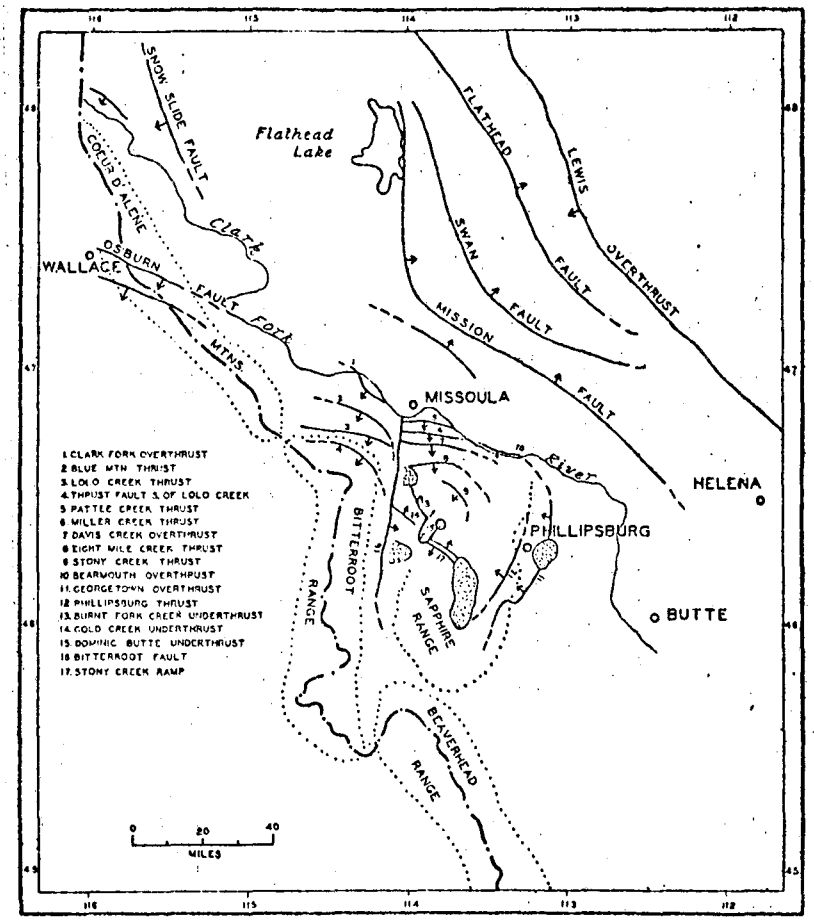


Figure 34. Index map showing major fault patterns of western Montana (Langton, 1935)

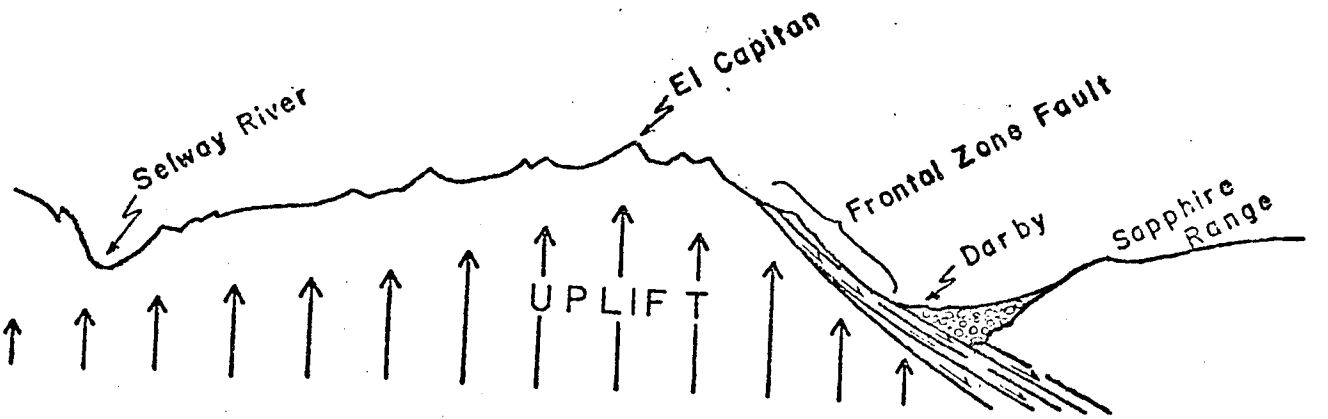


Figure 35. Hypothetical cross section: Bitterroot Mountains, Bitterroot Valley and the Sapphire Mountains, V.E.=8x (Winegar, 1974)

The Eastern Front of the Bitterroot Range Montana

Abstract

"The origin of the gneissic rocks on the eastern border of the Idaho batholith in the Bitterroot Range, near Hamilton, Mont., has long been in dispute. Lindgren regarded these rocks as the product of stresses related to a normal fault along the front of the range with an eastward dip of about 15°. He thought both the hanging wall and the footwall had moved, with a total displacement along the fault plane of at least 20,000 feet. The faulting was believed to have been so recent as to be a major factor in the present topography. Langton appears to accept the concept of faulting but to regard the gneissic rocks as formed much earlier from a granitic rock that was more silicic and older than the Idaho batholith.

Reexamination of the evidence in field and laboratory leads to the conclusion that the gneiss bordering the batholith in the Hamilton quadrangle more nearly resembles a sedimentary than an igneous rock, both in general aspect and in details of texture and composition. It is a stratified rock that has been recrystallized but has retained even the minor features of its original lamination. Feldspars and other minerals have been developed in it, probably as a result of permeation by emanations from the batholith during intrusion. The rocks from which the present gneissic mass developed differed among themselves in their details of composition. They are thought to have been components of the Ravalli group of the Belt series, with their characteristics now obscured by partial granitization. So many features of sedimentary rocks are present that the concept of the gneiss's derivation from an old granite is not regarded as acceptable." (Ross, 1950)

Surfical deposits fill the Bitterroot Valley and much of this material is lacustrine deposits from Lake Missoula. Several periods of glaciation are mentioned by all authors.

An extensive network of dikes (Tertiary ?) intrudes much of the Bitterroot Range and these are mostly quite felsic; in the range of quartz latite. These seem to intrude along the thrust faults which are associated with the batholith.

Because Section III, Subsection 9 is related to the general Bitterroot structural trend it is discused in this section.

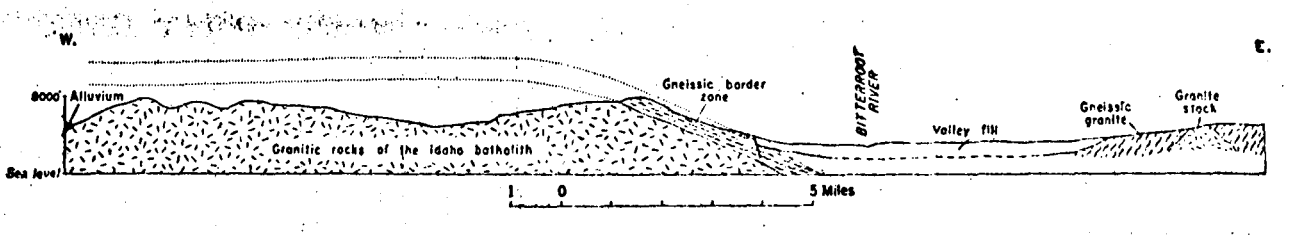


Figure 36. Hypothetical cross section: Bitterroot Mountains (Ross, 1950)

Subsection 1 - East Slope Bitterroot Range Subsection (Additional Comments)

The geology of a large part of Subsection 1 was compiled by Chase (1973). Chase noted that the metasediments within the area appeared to be equivalent to the Prichard Formation. Chase recognized five phases of folding in the metasediments. Several anorthosite bodies were also mapped. Chase's abstract follows.

Petrology of the Northeastern Border Zone of
the Idaho Batholith, Bitterroot Range, Montana

Abstract

"Pelitic, quartzofeldspathic, and calcareous sediments probably of the Belt Supergroup (Precambrian) have been regionally metamorphosed adjacent to the northeastern border of the Idaho batholith within the Bitterroot Range, Montana. The metasedimentary units comprise lithologically layered schists and gneisses, which contain concordant amphibolite pods and anorthosite bodies. Mineral assemblages are characteristic of sillimanite-potassium feldspar-grade or, locally , sillimanite-muscovite grade metamorphism.

Three phases of synmetamorphic penetrative deformation are recognized. Mica schistosity developed parallel to lithologic layering (probably bedding) during the first (F_1) phase. The second deformation (F_2) resulted in flexural slip folding of F_1 structures and development of mica crinkle lineation (paralleling F_2 fold axes) and incipient axial-plane schistosity. The third deformation (F_3) resulted in superposition of flexural slip folds on F_2 structures in quartzofeldspathic layers and concomitant development of crenulation foliation, axial-plane mica schistosity, and transposition structures, which nearly destroyed the F_2 form surface in pelitic layers. Calc-silicate layers were stretched into boudins during one of the penetrative deformation. Synkinematic igneous activity included the anatectic production of pegmatite veins and the emplacement of a tonalite stock in the quartzofeldspathic metasediments.

Emplacement of Idaho batholith quartz monzonite-granodiorite (texturally and chemically complex in nature) and associated apophyses, pegmatite, and aplite postdates regional metamorphism and structural deformation. Foliation in metasediments dipping under the batholith suggests a floor to the pluton in the area mapped, as does the topographic expression of the contact. Magmatic intrusion involving a complex interplay of forceful injection, magmatic stoping, and contact metasomatism is the probable emplacement mechanism. A macroscopic rotation (F_4) of structural subareas in the metasedimentary rocks adjacent to the batholith represents a possible effect of forceful

magmatic injection. Thermal recrystallization of quartz and micas in border-zone gneiss accompanied batholithic emplacement, but the mineral assemblages formed during regional metamorphism were in large part stable in the temperature-pressure field of igneous activity.

A segment of the Idaho batholith along the north-trending eastern front of the Bitterroot Range has been subjected to cataclastic flowage. This deformation (F_5) was dominantly penetrative, resulting in the superposition of quartz and mica subfabrics on the pluton." (Chase, 1973 p. 1)

Subsection 9 (Section III) - West Fork Bitterroot Streamcut Subsection

Subsection 9 is drained by the West Fork of the Bitterroot River. Winegar (1974) compiled and added to a geologic map which covers most of the subsection.

Five major lithologies are present in the area, 1) the Idaho batholith, 2) Precambrian, Belt equivalent metasediments (mostly quartzites and schists), 3) a zone of frontal gneisses and sheared granitics, 4) late stage quartz latites, and 5) volcanics.

Figure 37 shows the major tectonic setting. The following sequence of events is proposed by Winegar to explain this setting.

1. "Age of the frontal zone granite is uncertain. It is pre-deformation perhaps dating back to the Precambrian, but a better assumption is that it is an early Cretaceous intrusion -- perhaps contemporaneous with beginning uplift of the Bitterroots."

2. "Continued uplift causes the Sapphire plate to begin movement. Shear deformation forms in the frontal zone gneiss and frontal zone granite."

3. "After the shearing and movement of the Sapphire plate was complete or at least progressed further east, intrusion of a late Cretaceous granite (ki) occurred. This granite could indeed be Tertiary in age."

4. "The transecting normal faults develop, creating the en echelon contact pattern between the granite (Ki) and the gneiss (gn)."

5. "Erosion remained active and formed the topography we see today. Possibly during or shortly after this, fault movement was subsiding, and shallow intrusions of predominately quartz latite were emplaced. These intrusions often broke through to the surface and formed the volcanic deposits."

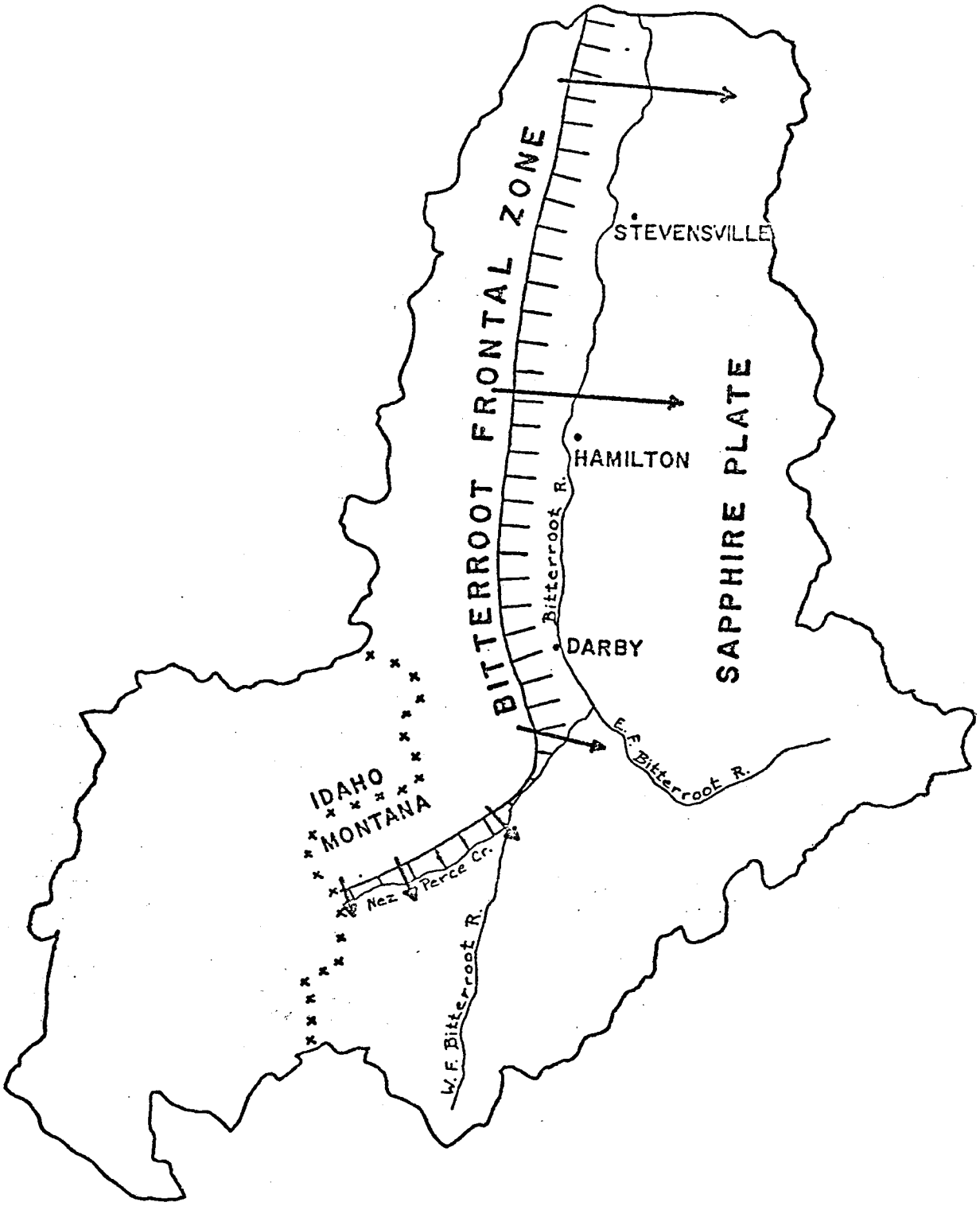


Figure 37. General tectonic map of the Bitterroot National Forest. Arrows show direction of movement of the Sapphire plate (Winegar, 1974)

> "The frontal zone granite may represent the original parent magma. The younger granite, the intrusive quartz latite and the volcanics could be later phases of crystallization of the same material. The younger rocks in general seem to be more felsic suggesting differentiation of the original granodioritic melt to silica and alkali-rich magmas." (Winegar, 1974)

Much of this area was glaciated in the higher elevations.

THE GEOLOGY OF SECTION VI

ST. JOE-CLEARWATER SECTION

Section VI is about evenly divided between the Hamilton and Wallace AMS maps with a little overlap onto the Spokane sheet. Most of the Section on the Wallace and Spokane maps is covered by very low grade metamorphosed Precambrian Belt units while the Hamilton sheet is covered by mostly higher grade metamorphic Belt units.

The western part of the Hamilton map area is drained by the North Fork of the Clearwater. Subsections on the Wallace map drain into the North Fork of the Clearwater, the St. Joe River system, or the South Fork of the Coeur d'Alene River.

Section VI is structurally complex but because of the extensive mining areas in the region much of the structure is fairly well understood. Kleinkopf, Harrison, and Zartman (1972) have presented a short summary of the area covered by Section VI as follows.

Aeromagnetic and Geologic Map of Part of Northwestern Montana and Northern Idaho

Introduction

"The area studied consists principally of Belt rocks that have been intruded by Precambrian gabbroic sills and by Mesozoic and Cenozoic plutons, sills, and dikes; most of these intrusives are intermediate to acidic in composition. Tertiary extrusive rocks are known in small parts of the area.

These rocks are exposed in several mountain ranges: from northeast to southwest, the Flathead Mountains, the Kootenai Mountains, the Cabinet Mountains, and the Coeur d'Alene and St. Joe Mountains of the Bitterroot Range. The Purcell Trench, which marks the eastern boundary of the Kootenai are mobile belt, crosses the northwestern part of the area and the Rocky Mountain Trench crosses the northeastern part. (See index to physiographic names and mining districts.)

The famous Coeur d'Alene mining district, as well as several other districts that also have histories of sizeable silver-lead-zinc production, is located in the area. Among the other districts are the Lakeview, Clark Fork, and Talache mining areas of Idaho, and the Mineral County, Hog Haven, Cabinet, Libby, Troy, and Blue Creek mining areas of Montana.

This map relates geology, aeromagnetic data, and lead isotope data to the vein-type mineralization of the area. Although the aeromagnetic data are complete, both the geologic and the isotopic studies are still in progress. Partly because of the preliminary nature of the geologic

and lead isotope data, no quantitative analysis of the magnetic data has been undertaken. All data herein are being released in preliminary form before completion of the geologic and isotopic studies, which will take several more years of work, because the pattern of association of certain ores with certain intrusives or with positive aeromagnetic anomalies suggests that the aeromagnetic data can be used as a prospecting tool in some areas not yet thoroughly explored.

Kleinkopf was responsible for interpretation of the aeromagnetic data, which have been previously released for public use through an open-file report and publications of the U. S. Geological Survey (U. S. Geol. Survey, 1969a, 1969b-k; Meuschke and others, 1962). Harrison compiled the geologic data from many sources (see index map), and he made only minor changes from the published maps where it was essential to join various segments. Areas inadequately mapped for compilation at 1:250,000 were left blank on the map except for a few important faults or contacts, which are shown in dashed or dotted lines. Zartman was responsible for interpretation of the isotopic data. Inferences as to the significance of the three types of data are preliminary but are mutual among the three authors."

General Geology

"Rocks of the mapped area belong dominantly to the Precambrian Belt Supergroup, a stratigraphic sequence at least 42,000 feet thick (the base is not exposed) of argillite, siltite, quartzite, and, in minor amounts, carbonate rocks. The oldest unit is the Prichard Formation, which consists characteristically of laminated black and gray pyritic or pyrrhotitic rusty-weathering argillite, gray argillite and siltite, and, in the lower part, gray impure quartzite. Next younger is the Ravalli Group, which consists of three units. The lower unit (the Burke Formation) is gray to purple-streaked magnetite-speckled siltite interbedded with lesser amounts of green to purple argillite. The middle unit of the Ravalli is the Ravett Formation, which contains characteristic beds of white, tan, or purple-streaked blocky crossbedded quartzite. This unit thins from about 2,000 feet in the west to a few hundred feet in the northeast. The upper unit of the Ravalli is the St. Regis Formation in the southern and western parts of the area and the Grinnell Argillite in the northeastern part. The St. Regis is an interbedded carbonate-bearing sequence of purple argillite and siltite; it thins from southwest to northeast and has not been identified cast of Libby, Mont. The Grinnell is dominantly purple argillite, but contains a few green argillite and light-colored siltite or quartzite beds, some of which are carbonate bearing. Overlying the Ravalli

Group are the two carbonate-bearing formations, the Wallace Formation and its stratigraphic equivalent, the Helena Formation. The Wallace, which is a thick sequence of carbonate-bearing laminated black and white argillite, green argillaceous siltite, and sparse beds of limestone or dolomite, is exposed in most of the area. In the northeastern part of the area, this stratigraphic interval is represented by the Helena Formation, which contains some of the black and white argillite and green siltite beds similar to the Wallace Formation but which has thicker and more abundant layers of limestone and dolomite. Overlying the Wallace Formation are rocks of the Missoula Group. These rocks are dominantly red and green argillites, siltites, and quartzites that contain a few dolomite beds or dolomitic layers. The Missoula Group thins from about 16,000 feet in the southeast to about 4,000 feet in the Pend Oreille Lake area; part of the thinning is due to erosion that occurred before deposition of Cambrian sediments. All formations in the Missoula Group are thicker in the southeast, and subtle unconformities within the group appear to have cut out or thinned some formations; an example is the conspicuous red feldspathic quartzite (Lupine Quartzite) that is about 3,000 feet thick near Superior, Mont., but thins to about 700 feet near Pend Oreille Lake where it forms the upper member of the Striped Peak Formation.

Metamorphism of the Belt rocks increases from east to west and with depth in the stratigraphic section. Rocks of the Burke Formation and those immediately below the Burke are in the biotite zone of the greenschist facies. In the northwestern and southwestern parts of the area, Belt rocks are now schists and gneisses.

Belt rocks are overlain unconformably in a few places by Paleozoic sedimentary rocks. The basal unit (the Flathead Quartzite or its equivalents) is a quartzite.

Structurally, the area is basically a Precambrian anticlinorium of north-northwest-trending folds and faults. This simple structure was modified in Precambrian time by right-lateral offset on the ancestral Osburn and Hope fault zones that trend west-northwest across the area. It was further complicated by late Cenozoic block faulting that created basin-and-range structure, which now dominates most of the area north of the Osburn fault zone. The Libby Trough, which extends north-northwest across the central part of the area, contains dropped blocks of Paleozoic rocks. Preservation of the Paleozoic rocks is a conspicuous result of this block faulting, and the structural trough is probably more continuous than shown on the compilation. South of the Osburn fault zone, additional folds and faults, not yet completely understood but apparently of both Precambrian and younger ages, surround the Idaho batholith: the northern edge of the

batholith is a few miles south of the mapped area. Late Cretaceous folding and thrusting at the eastern edge of the Kootenai are mobile belt (west of the Purcell Trench) complicate the structural pattern in the northwest corner of the mapped area. Renewed movement of the Hope and Osburn fault systems in Tertiary time is also reasonably well documented.

Various types of igneous rocks have intruded or have been extruded onto the Belt rocks. Precambrian sills of quartz diorite or gabbro are particularly abundant in the northwestern part of the area. Many sills have been metamorphosed and altered, but a few retain a pronounced igneous texture and a moderate magnetite content. Cretaceous intrusive rocks show a wide range in composition and form. Most are plutons of granodiorite or quartz monzonite that have a contact metamorphic aureole a few thousand feet wide. One syenite pluton that has the most pronounced magnetic expression in the area is exposed on Vermiculite Mountain in the north-central part of the area. Tertiary intrusives are largely dikes and sills; some of them are porphyritic.

Relatively small exposures of Tertiary volcanic rocks include the isolated latitic and andesitic rocks in the east-central part of the area and the eastern edge of the vast Columbia River Basalt flows in the southwestern part of the area." (Kleinkopf, Harrison, and Zartman, 1972)

Subsection 1 - North Fork Clearwater Rejuvenated Canyon Lands Subsection

Subsection 1 includes the eastern part of the area mapped by Hietanen (1968). The major drainage in the subsection is the North Fork of the Clearwater. Dominant lithologies are Prichard equivalent rocks in the northwest, Wallace equivalent rock in the northeast and south and Idaho batholith (quartz monzonite, Ki3c) in the central part. Larsen and Schmidt's (1958) road log shows a band of quartz diorite (Ki2a) which is covered to the north by undifferentiated Belt rocks (mpCl). The quartz diorite is most likely the roof facies of the batholith. The Hamilton linear map shows a strong northwest linear trend (which is related to the Trans-Idaho Discontinuity) through much of the subsection.

Subsection 2 - Deception-Hoodoo Basin and Uplands Subsection

Very little data is available on Subsection 2. The major drainage in the area are tributaries to the North Fork of the Clearwater. The northern part of the subsection was mapped by Anderson (1930) as Precambrian Belt and Norman Day (pers. comn.) has substantiated this although he noted that several outcrops of Idaho batholith occur in the area. The southern portion was mapped by Greenwood and Morrison (1973) as quartz monzonite-granodiorite of the batholith. Larsen and Schmidt's (1958) road log through the southern part of the

subsection recorded a long stretch of undifferentiated Belt rocks but no other particulars were given. Because Subsection 2 borders the western edge of the Bitterroot Mountains, it should be similar to Section IV, Subsection 5, that is, foot hill valleys filled with glacial outwash. The whole area just south of Deception Point to Kelly's Thumb and Tobaggan Hill has a topographic expression which is more indicative of Idaho batholith than Belt metasediments which is possible as Anderson (1930) mapped this area on a broad reconnaissance basis.

Subsection 3 - Bitterroot Strongly Glaciated Land Subsection

Subsection 3 contains the northern portions of the Bitterroot Mountains which marks the border between Idaho and Montana. The formation of the Bitterroots has been discussed in Section V. There is little data on this subsection. According to Reid (pers. comm.), the zeolite/greenschist facies border on the Wallace map runs just south of the Bitterroots and then west to the St. Joe River and continues along the river onto the Spokane map (see double line on maps).

Subsecton 4 - Lolo Strongly Dissected Land Subsection

Subsection 4 is drained by northward flowing tributaries of the Clark Fork. There is very little known about this area but on the basis of Langton's (1935) work in the eastern part of 4 Nold's (1968) thesis to the south in Section IV, Subsection 5, Kleinkopf, Harrison, and Zartman's (1972) work to the northwest and Anderson's (1930) reconnaissance geology to the west it appears that Subsection 4 should be mostly unmetamorphosed Belt. The major structural trend through this region should be northwest and this is the way that most of the streams flow implying fault control of the tributaries.

Subsection 5 - Wade Peak Subsection

Subsection 5 parallels the St. Regis River and the Clark Fork to the southeast. Five is very similar to Subsection 10. The northwestern part of the subsection was mapped by Wallace and Hosterman (1956) while the southeastern part was mapped by Campbell (1960).

Lithologies in the area are the usual Belt sequence. The major structure in the subsection is the Osborn fault (just north of the section boundary). Wallace and Hosterman summarized the structural setting as follows.

Structural Geology

General Setting

"The region has been subjected to intense folding and faulting. The major structural feature is the Osburn fault zone which approximately bisects the mapped region along a line trending N. 65° - 70° W. All other major faults and fold axes roughly parallel this trend, although in gross pattern they trend slightly more north of west and appear to be truncated at a very low angle by the Osburn fault zone.

Deformation is much more intense in the structural block north of the Osburn fault zone than in the block to the south. On the north side of the fault zone overturned folds are the common structure, but on the south side overturning has occurred in only a few places. Most of the western half of the area south of the Osburn fault zone is underlain by a faulted homocline dipping south. The eastern half of the homocline is complicated by two eastward-plunging anticlines, and the north flank of each is truncated by a fault. South of the Osburn fault zone near the extreme western edge of the map are two overturned folds, and at the extreme eastern edge of the map the Wallace formation has been intricately folded. This alternation of intensely folded and relatively unfolded blocks south of the Osburn fault zone is typical of the area between Kellogg, Idaho, and St. Regis, Mont.

The Osburn fault zone itself is a complex of anastomosing faults, shears, and shattered zones ranging in composite width from 1,000 feet to 2 miles. A broad elongate topographic trough has been eroded in the soft broken rock along the general line of the Osburn fault zone. Pleistocene and Recent deposits occupy this trough and almost everywhere mask the fault zone, but gouge and intensely sheared rock are exposed in several places." (Wallace and Hosterman p. 588)

Glacial deposits of several periods are found on the ridges in the area as are lacustrine deposits of Glacial Lake Missoula. Campbell (1960) discussed the lake as follows.

Glacial Lake Missoula Deposits

"The Clark Fork valley in the area of this report was flooded during part of the Pleistocene epoch by an extensive lake. The history of this glacial Lake Missoula has been discussed by Pardee (1910; 1942), Alden (1953, p. 154-160), and Bretz and others (1956). The water level of the lake was reported to have reached a maximum altitude of 4,150 to 4,200 feet. It occupied an area of about 3,000 square miles and had a total volume of more than 500 cubic miles. The lake may have filled and at least partly emptied several times during its history.

Several poorly defined wave-cut benches mark the elevations of stillstands of the lake level at several places on the walls of the Clark Fork valley. The best examples of these benches occur on the east-facing valley wall in sec. 7, T. 17 N., R. 27 W., but even there they are obvious only from a great distance when the sun is in a favorable position to cast accentuating shadows on them.

Fine silt and clay deposits that probably were deposited in this lake were found in several localities on some of the terrace levels. Patches of erratic cobbles and

> boulders, generally of rocks from the Missoula group were found at several localities at an altitude of 4,000 feet or more and probably have been ice rafted on the lake. A large angular erratic boulder of the Wallace formation is resting on a small terrace on the spur between Dry and Thompson Creeks where it is difficult to attribute its position to anything other than ice rafting." (Campbell, 1960 p. 575-576)

Subsection 6 - Upper St. Joe Strongly Dissected Lands Subsection

Subsection 6 includes the headwaters of the St. Joe River (Hamilton AMS map) and the upper part of the St. Joe River (Wallace AMS map). Six is in the hinge zone where north-south structure has turned to north-west structure. The major structural feature in the area is the St. Joe fault which like the Osburn Fault is a strike-slip fault with left lateral displacement which may total 20 miles. The St. Joe River is controlled by this structure as it follows the regional structural trend.

As has been noted several times before the zeolite/greenschist facies border closely parallels the St. Joe River. The metamorphic rocks in the subsection are low grade and are members of the Wallace Formation (mpCl3).

Subsection 7 - Sisters Basin and Uplands Subsection

Subsection 7 is included in the general geologic map of Hietanen (1962). The northern half of 7 drains into the St. Joe and the southern half drains into the Little North Fork of the Clearwater River.

Like Subsections 5, 6, 8, and 10, 7 follows the northwest tectonic trend in the region. The general trend of these subsections may be controlled by 1) vertical movement on faults which parallel these areas, 2) structural anticlines and synclines which parallel the large strike-slip faults, or 3) a combination of folding and faulting.

The northern part of the subsection contains a good sized outlier of the Idaho batholith. A small piece of one of the anorthosite bodies outcrops to the south. The rest of the area is covered with Prichard Formation (mpCll) and Ravelli Group (mpCl2) metasediments.

Subsection 8 - Marble Mountain Strongly Dissected Uplands Subsection

Subsection 8 is drained by Marble Creek which flows north to the St. Joe River. The subsection is about equally divided between the Wallace and Spokane AMS maps. The western border of 8 coincides with a major north-south fault zone (see Spokane AMS map). Just west of this fault zone is a double dashed line which represents the approminate boundary between the zeolite and greenschist metamorphic facies. Eight is in the greenschist zone. The eastern border of 8 is also marked by a north-south fault. The southern part of the area extends onto the Hamilton map where a part of an

anorthosite body is located. Metamorphic grade is in the Staurolite Kyanite range (see Figure 32). Belt rocks in 8 consist of the Prichard (mpC11) with some Wallace Formation (mpC13).

Subsection 9 - Baby Grand Low Relief Uplands Subsection

Subsection 9 is located in the southeast corner of the Spokane AMS map. Major drainages in the area are the St. Joe River (north), and the St. Maries River (central), (a major tributary of the St. Joe). The double dashed line in the center of 9 is the zeolite/greebschist boundary. Belt rocks west of the line are poorly metamorphosed while east of the line rocks are visibly recrystallized. All four major Belt units are present in the area and a large part of the central region of the subsection is covered with undifferentiated Columbia River basalt (Te_1C).

Major structures in the subsection are the St. Joe fault (parallels river) and probable north/south faults which are related to the development of the Purcell trench to the north. Griggs (1973) described the Belt Supergroup and the general geology of the Spokane AMS map as follows:

The Belt Supergroup in the Spokane Quadrangle

"Rocks of the Belt Supergroup of Precambrian age form the bedrock under most of the east half of the 1^{o} by 2^{o} Spokane quadrangle and are exposed in scattered hills and ridges that protrude above the basalt plateau on to the west. These form the westernmost exposures of the Belt over a north-south distance of about 70 miles. All formations of the western section from Prichard up through Libby are exposed, totaling about 40,000 feet with neither the top nor bottom exposed. Facies changes are noticeable in some units, but the overall similarity of each formation throughout the quadrangle and to correlative units on to the east is more remarkable. All of the Belt rocks have had a low grade regional metamorphism imposed upon them; in the lower part of the section it reaches the green schist facies (biotite zone). Two areas are underlain by highly metamorphosed Belt rocks; one is in the southeast corner of the map and the other is in the north central part. That to the southeast is part of the broad progressively metamorphosed aureole around the Idaho batholith. The other bounds part of the east and southeast margin of the Loon Lake batholith. These large multi-pluton complexes were emplaced during a period spanning much of Mesozoic and into early Tertiary time; this was also the time of the more intense metamorphism. Some of the resulting schist and gneiss is well up into the amphibolite facies. Through warping and faulting some of the high-grade rocks have been elevated to where they are locally in direct contact with normal Belt rocks. North- to west-trending folds and a complex of faults have imposed an intricate structural pattern on the Belt rocks. A zone about 25 miles wide of west-northwest

trend (the Lewis and Clark Line) extending across the east central part of the quadrangle is characterized by a net of subparallel faults. The Osburn and St. Joe faults of major right-lateral strike slip dominate the zone and many other faults of like but lesser movement are present. Faults on which the movement has been mostly dip-slip also are numerous. To the north and south of this zone block faulting is the characteristic patern. This structural deformation has resulted from a number of periods of activity beginning in Precambrian, but most came about during the intermittent orogeny that continued through much of Mesozoic and early Tertiary time." (Griggs, 1973)

Subsection 10 - St. Joe Thrust Fault Lands Subsection

Subsection 10 includes part of the Coeru d'Alene mining district and has been mapped in detail by Hobbs et. al., (1965). The main drainages in the subsection are the St. Regis River system on the north and east and the St. Joe River to the south and west. Maximum relief in the area is 3,000 - 4,000 feet which results in rugged terrain. The upper ridge crests reach 6,000 - 7,000 feet.

The regional geologic setting of this area revolves around the Precambrian Belt geosyncline which extended from Canada through western Montana. Upwarp of this mass of Belt sediments was probably eroded away by middle Cambrian. This upwarp was gentle with little deformation. By middle Cambrian, the upwarp was worn down to low relief and was covered with Cambrian sediments. Emplacement of the Idaho batholith occurred in Late Cretaceous times.

The regional structural history of the area is summarized by Hobbs et. al., (1965) as follows.

"The structural pattern of the region surrounding the Coeur d'Alene district is characterized by two outstanding and significant features: (a) the gently deformed condition of the rocks over large areas--intensely folded only locally, and (b) a series of northwest-trending major faults of large lateral displacement. The generally uncomplicated structure of the Belt Series rocks has been recognized and speculated upon from the earliest work in the region. There are, of course, numerous local steep and even overturned folds, but generally the rocks occur in great broad, open folds in which the dips are usually less than 45°, and in large areas where the rocks are mainly flat. Fold axes generally trend north to northwest in that part of northern Idaho and western Montana north of the general latitude of the Coeur d'Alene district. In the vicinity of this latitude and south to the Idaho batholith, the fold axes trend more westward.

The rocks in and near the Coeur d'Alene district are cut by a complex of faults that have complicated histories,

but the whole pattern of this deformation is dominated by three great transverse faults of fault zones, each of which is marked by major drainage lines: (a) the St. Joe fault in the valley of the St. Joe River in Idaho, (b) the Osburn fault and its related structural features in the valleys of the South Fork of the Coeur d'Alene River in Idaho and the St. Regis River and parts of the Clark Fork in Montana, and (c) the Hope fault in the lower valley of the Clark Fork in Idaho and Montana. Of these great faults, the Osburn fault is the most profound and extensive, and forms a large part of the pronounced linear feature that Billingsley and Locke (1939, p.36) called the Lewis and Clark line.

From this picture the Coeur d'Alene district can be seen to lie in the central part of a late Precambrian geosyncline of north-northwesterly trend. Furthermore, the district is located at the intersection of a major anticlinal arch of these geosynclinal sediments and a major structural zone of weakness, the Lewis and Clark line, which lies nearly athwart this geosynclinal basin. The downbuckling that produced the eastern prong of the Belt basin, the deformation in the Coeur d'Alene district prior to the emplacement of the 1,250 year-old uraninite veins, and the movement along the Osburn fault system all appear to be reflections of sporadic activity along the Lewis and Clark line. It seems likely that the Coeur d'Alene district is within an area that has been in a positive, emergent position since early in the Paleozoic era. Large masses of igneous rock were intruded into the Belt Series in the western part of the Belt basin in Mesozoic time. Then the mineral deposits of the main period of the Coeur d'Alene district were formed near one of the satellite intrusions within a structural knot in which the rocks had been highly fractured." (Hobbs p. 12)

Lithologies within Subsection 10 are confined to the Belt Supergroup rocks which have been described elsewhere. Several small stocks of quartz monzonite intrude the Belt rocks as do dikes of aplite, diorite, dacite, and lamprophyres.

The geomorphologic history of the area is summarized as follows.

Geomorphology and Surficial Deposits

"The Coeur d'Alene district lies within the werstern part of the Northern Rocky Mountains physiographic province (Fenneman, 1931, p. 186), which occupies much of central and northern Idaho and adjacent parts of Montana. This province is a continuous mountain mass wherein individually distinct ranges are nonexistent. The streams usually are deeply entrenched in steep-walled narrow valleys. In contrast, the eastern part of the province, which occupies much of western Montana and parts of south central Idaho,

consists of definite elongate ranges separated by intermontane basins.

Within the Coeur d'Alene district the upland parts consist of long, sinuous ridges that commonly maintain a fairly uniform altitude, although the average height gradually decreases westward from the Bitterroot divide. At a few places the usual narrow crests broaden into small flats. A view of the horizon from almost any high vantage point shows a pronounced accordance among the tops of the ridges in and around the district; a few scattered peaks stand out only several hundred feet above the general level. The valley of the South Fork of the Coeur d'Alene River westward from Wallace is the only relatively large area of low flatland; at places the valley floor is almost a mile across. Most of the tributary valleys are steep walled and V-shaped, although the floors of some widen appreciably in their lower courses.

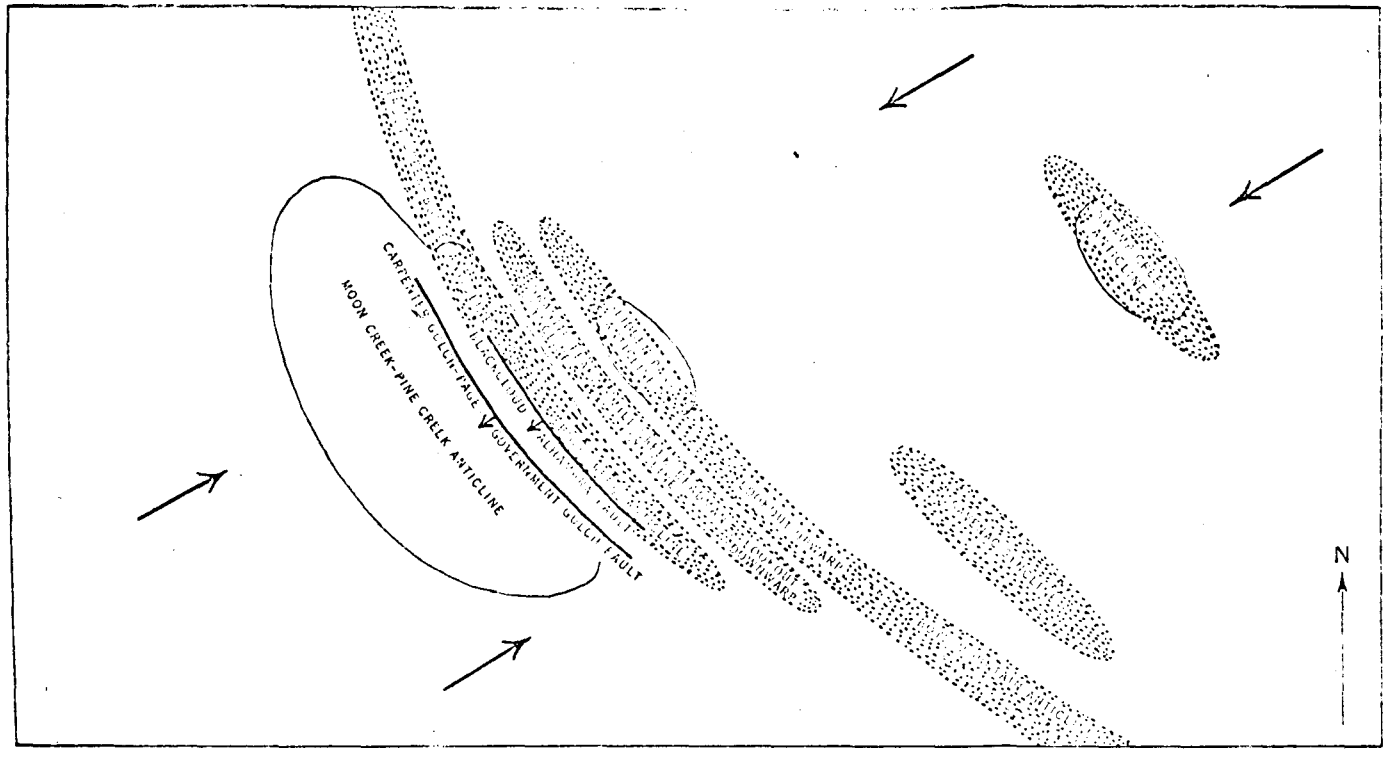
The present landscape in the Coeur d'Alene Mountains appears to have resulted from the dissection of a mature land surface. The geomorphic history, however, is not simple; on the contrary, it is multicyclic. At least two principal periods of aggradation and numerous subperiods of short duration are interspersed through the general history of downcutting. All this is evident from the flats and notches, some of which are gravel floored, on the subsidiary ridges below the accordant summits of the main high crests. The two principal periods of aggradation can be related to outside influences: the first to the damming of the ancient Coeur d'Alene River in middle Tertiary time by Columbia River Basalt flows which then built up the plateau to the west, and the second to the damming of the main stream valley at the Purcell Trench by lobes of continental ice and the material they deposited. A continuing but interrupted uplift of this mountainous area in comparison to the country to the west is indicated for all time since the beginning of the dissection of the summit-capping mature land surface. In addition, alpine glaciers in Pleistocene time modified the landform by the cutting of cirques in the higher ridges and deposition of debris in the valleys below.

The assignment of definite ages to many of the events in this geomorphic history is impossible, particularly when only the evidence found within the district is considered. The geologic event of most recent date within the mapped area which can be used to limit the beginning of the general cycle of erosion was the intrusion of the monzonitic stocks in Cretaceous time. Thus, in order to date even tentatively post-Cretaceous events in the area, much evidence gained from investigations outside the district

has been drawn upon.

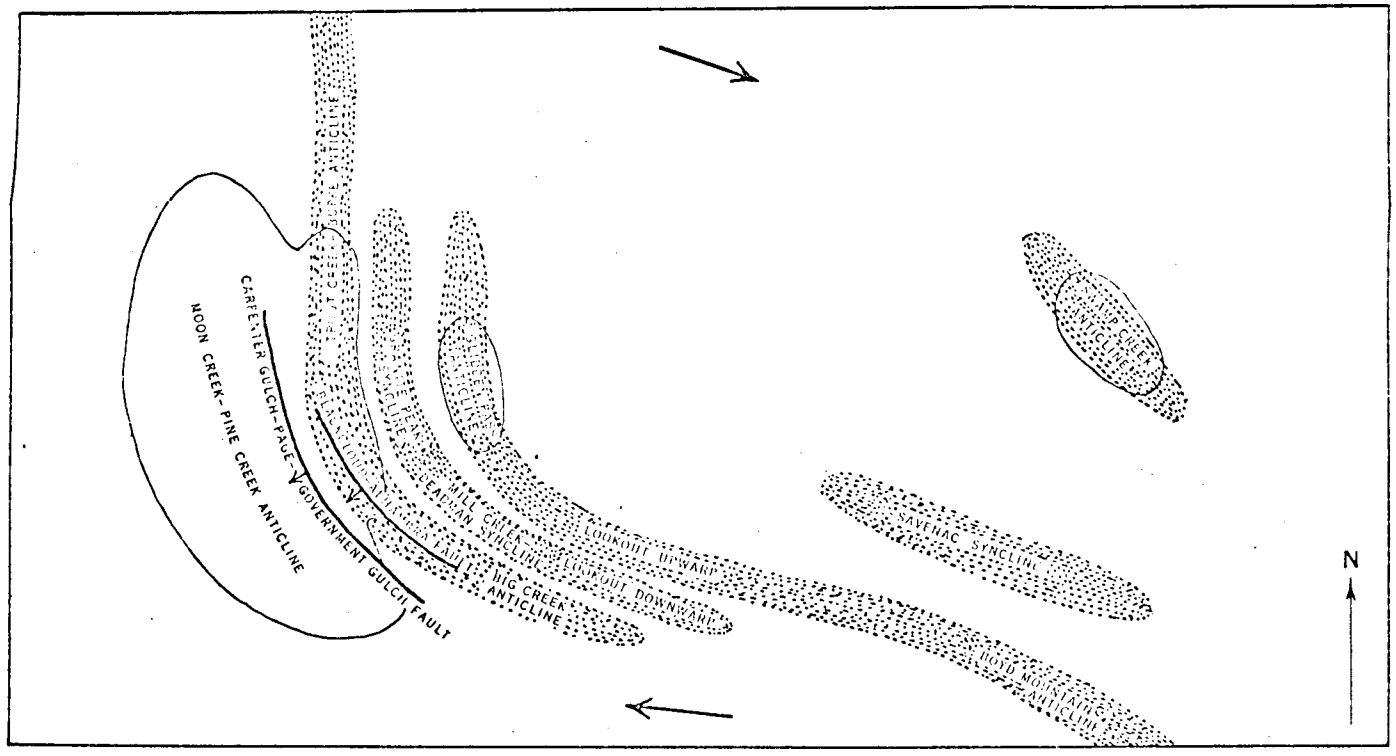
> Structural features have only indirectly controlled the surface configuration in the Coeur d'Alene district and the region that surrounds it. Deformation preceded the sculpturing of the present landscape sufficiently long ago that none of the original irregularities are reflected by the topography today; no fault scarp or fold can be traced by its surface expression. However, major faults such as the St. Joe to the south, the Osburn fault within the district, and the Hope to the north have undoubtedly controlled the position of major drainage channels for at least part of their courses. The valley of the South Fork of the Coeur d'Alene River closely follows the trace of the Osburn fault throughout the district. The superposition is not perfect, but is too close to be fortuitous. The courses of parts of Ninemile and Beaver Creeks and the low divide between them are apparently the direct result of the ease of erosion along the Dobson Pass fault. The Placer Creek fault lies athwart the drainage pattern, but even so several tributary gulches follow closely along its trace, and ridges usually contain a saddle where the fault crosses. Many additional faults have similar surface expression; for many other, however, little or no similar evidence can be found." (Hobbs p. 63-64)

The regional structure picture presented earlier mentions that deformation, overall in the Belt geosyncline was minor with local major deformation. Subsection 10 is such an area of strong deformation. Most of this deformation is related to the Osborn fault system (trends west northwest) which has distorted the broad folds of the geosynclinal basin which trends north-south. Figure 38A through F show the generalized twisted evolution of the Coeur d'Alene area. A more complete discussion of the structure in this region is presented by Cady (1974).

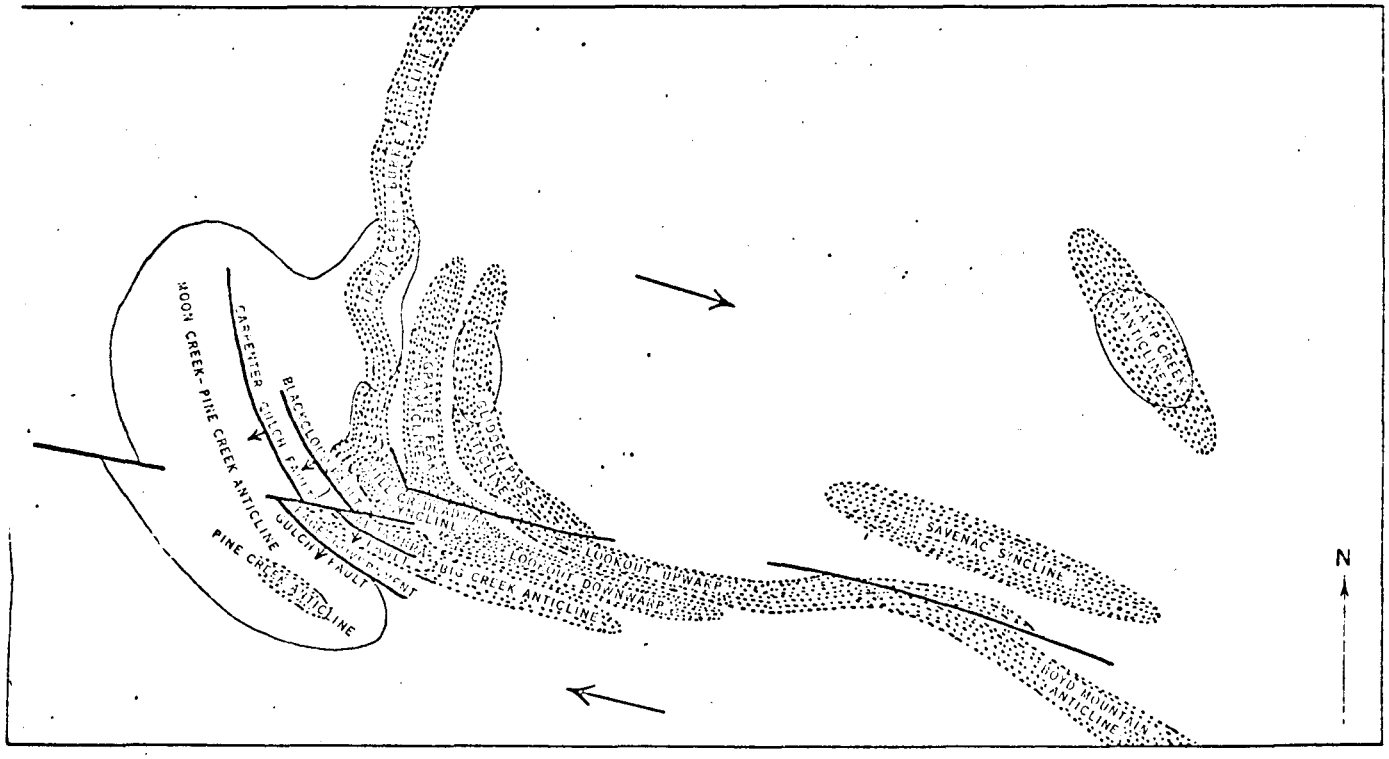


A, In the first stage of deformation the principal folds are formed by stresses oriented northeast and southwest. In the tighter segments the folds are overturned where reverse faults that strike northwest and dip southwest were formed. A large upwarp, the Moon Creek-Pine Creek anticline, lies elongated parallel to and west of the folds, which together with the Glidden Pass and Swamp Creek anticlines outlines a crude culmination of folding that transects the fold axes.

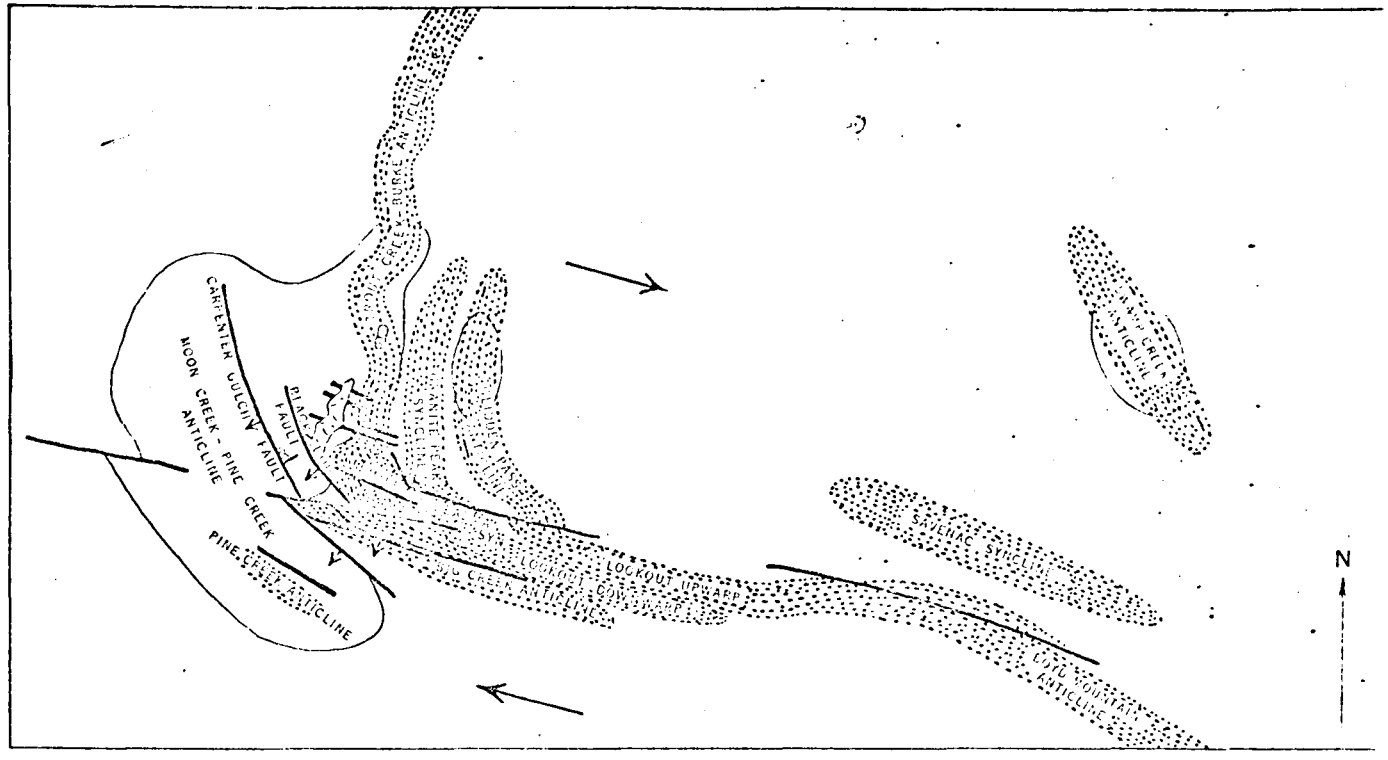
Figure 38 - Structural development of Coeur d'Alene District - Series A-F (After Hobbs et. al., 1965)



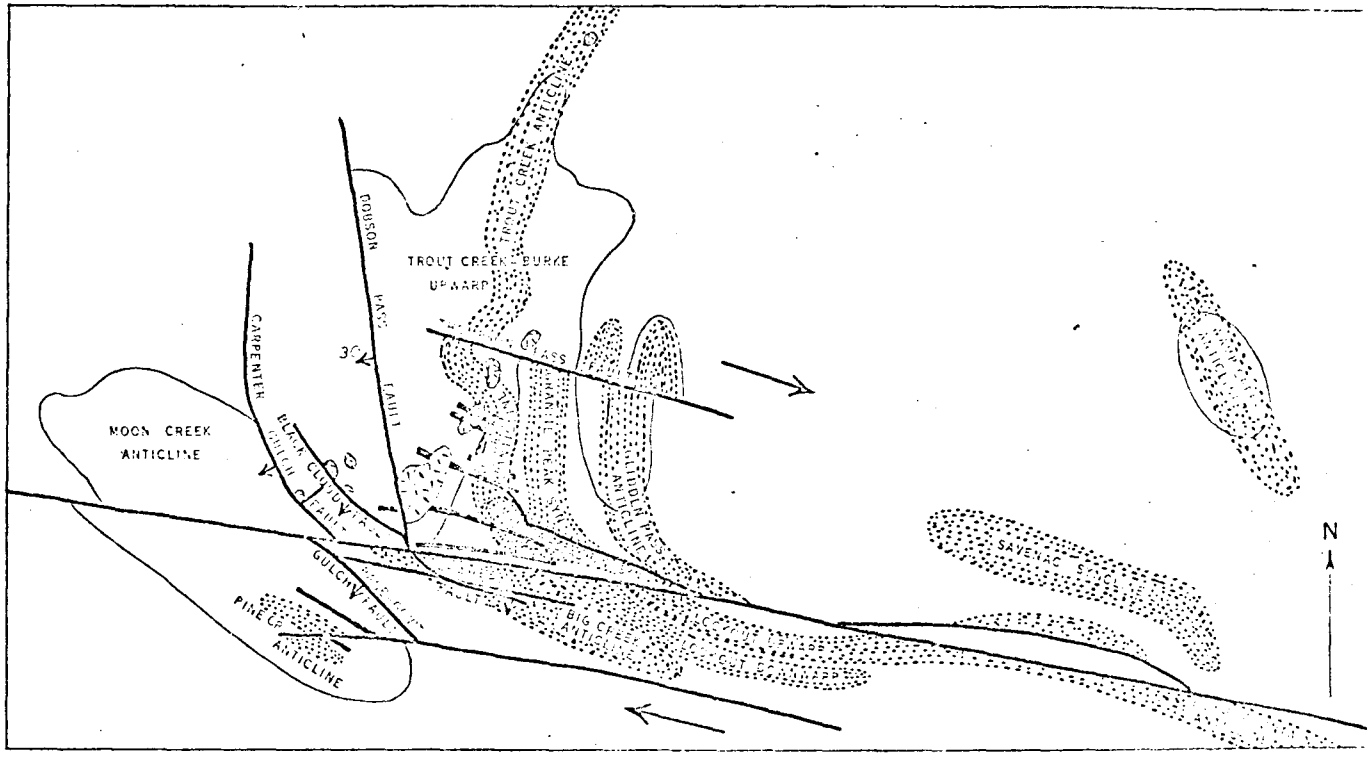
B, A major reorientation of the stress system begins to bow the axes of the folds. These stresses, rotational in plan view, drag the southern part of the region westward relative to the northern part. Stresses were probably distributed from below the crust.



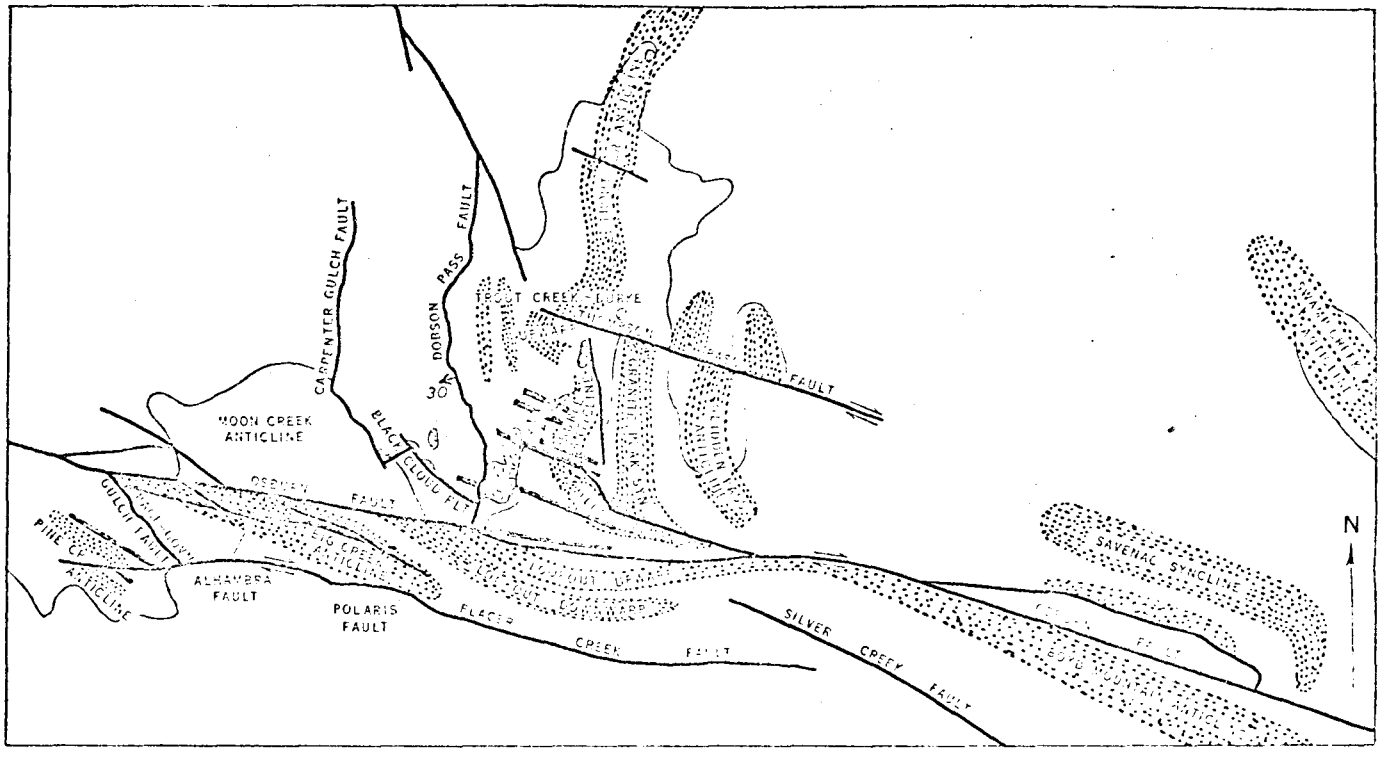
C, The rotational stress system has produced tight crenulations in fold axes and incipient strike-slip faults. A structural knot is thus produced in what is to be the heart of the Coeur d'Alene district. The Mill Creek and Deadman synclines have separated from the Granite Peak syncline and wrapped around the truncated end. The north flank of the Lookout-Boyd Mountain anticline is sliced off by an early segment of the Osburn fault which later becomes the north branch of the Osburn fault in Mineral County, Mont.



D, Monzonitic stocks have intruded the structural knot and are elongate almost parallel to folds. The major veins have formed probably as a late phase of the igneous activity. Mineralizing solutions have passed into the crustal rocks through several parallel feeders in the basement that are linear and that possibly originated as incipient strike-slip fractures opened by clockwise rotational stresses. Veins in the crust above these feeders occupy fractures of several types and orientations, but most are subparallel to the feeder zones.



E, Rotation has continued and strike-slip shear zones have become continuous fault zones. These are more throughgoing than previous faults and more nearly represent a direct release of fundamental deep-seated stresses than do the earlier faults, which were more likely produced in the upper skin of the crust by secondary stresses indirectly transmitted from block to block and which were subject to the vagaries of local discontinuities. With the formation of such throughgoing faults, it is possible that most of the deep-seated stresses are accommodated by displacement along these and only a few other faults. The Thompson Pass fault has begun to offset the folds, and the Placer Creek fault has begun to offset the Pine Creek anticline and Yreka vein system. The Osburn fault has begun to offset the major folds and separate the northern vein systems from the southern ones. The original Moon Creek-Pine Creek anticline has been cut and partly offset.



F, Present position of structural elements. The Moon Creek-Pine Creek anticline and the two thrust-fault zones were displaced by the Osburn fault. Although this step in the displacement of the block west of the Dobson Pass fault from the block south of the Osburn fault is shown in *F* separately from *E*, the displacements shown in both probably took place, at least in part, concurrently. Some of the earliest faults to be activated as strike-slip faults are flexed when later strike-slip stresses are accommodated along slightly different planes. The Placer Creek fault and branches of the Osburn fault in Montana thus have somewhat irregular trends. The east end of the Savenac syncline and the early-formed north branch of the Osburn faults are sharply bent. The Polaris fault takes up strike slip after the Placer Creek fault has buckled.

THE GEOLOGY OF SECTION VII

OPEN NORTHERN ROCKIES SECTION

Section IX contains part of the Salmon River Mountains to the west, the Lemhi Range to the south, the Beaverhead Range to the East and the Bitterroot Range to the north. The central portion of the Section is covered by the Salmon Basin which connects to the Lemhi Valley to the south. Anderson (1956, 1957, 1958, 1961) mapped all of Section IX which is covered in this report.

Elevations in the Salmon area range from 4,000 feet near Salmon to over 9,000 feet in the Salmon River Mountains to the west and well over 10,000 feet in the Beaverhead Range (Continental Divide). The Lemhi Range terminates in the southern part of the area but extends from this termination for over 100 miles south to the Snake River Plain.

Major drainages in the subsection include the Salmon River and the Lemhi River which join at the town of Salmon. Ruppel (1967) discussed Tertiary changes in the flow directions of the Lemhi River and the abstract of this paper follows (Figure 39).

Abstract

"The Lemhi River, a northwest-trending river in east-central Idaho, apparently flowed to the southeast until late Pliocene or early Pleistocene time, and was then reversed by the rise of the Gilmore Summit. The river and its parallel--and probably also reversed--sister streams, the Pahsimeroi River and Warm Spring Creek, are now tributaries to the Salmon River, but the Salmon could not have existed in anything resembling its present form until after the reversal. The Big Hole River, in adjacent southwestern Montana, may also have drained to south at one time, which suggests that fossil placer deposits of gold might be found in the Big Hole drainage area by analysis of late Cenozoic tilting and faulting and reconstruction of late Pliocene and early Pleistocene drainage systems." (Ruppel, 1967)

The reversal of the Lemhi and Pahsimeroi Rivers was probably related to the Snake River downwarp (see Section I discussion).

The major lithologic unit in the section is a thick sequence of Precambrian quartzites which form the major Mountain Ranges in the region. Challis volcanics, Idaho batholith granitics, some Tertiary dikes and a variety of fluvial and glacial surficial deposits cover or intrude large areas of the quartzites.

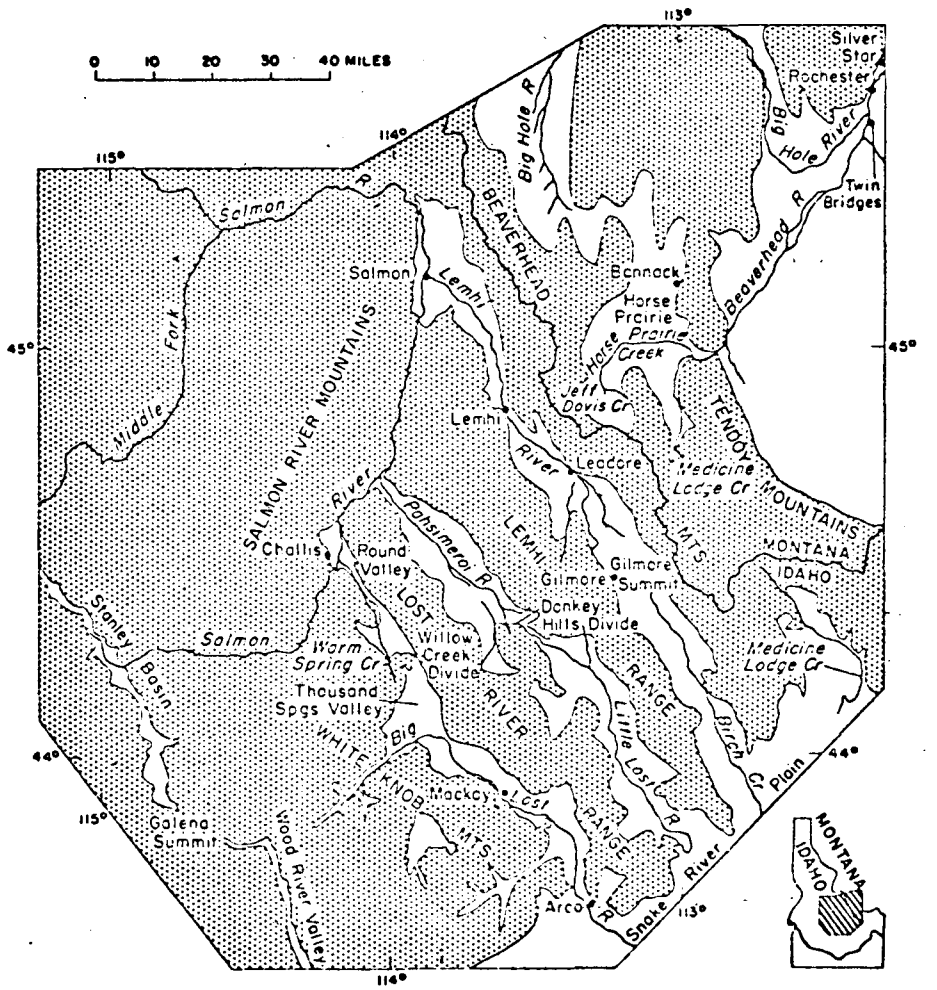


Figure 39. Drainage location maps: East-central Idaho to adjacent Montana (Ruppel, 1967)

Subsection 1 - Hughes Point-Saddle Mountain Subsection

Subsection 2 - Gibbonsville Subsection

There is little information available on Subsections 1 and 2, but the road log of Larsen and Schmidt (1958) and some road log observations by Ross (1962), infer that most, if not all, of 1 is covered by undifferentiated Precambrian metasediments (mpCl). Foster (pers. comm.) noted that most of the Gibbonsville Subsection was covered by Precambrian quartzites and schists.

Subsection 3 - Salmon-Bitterroot Range Subsection

Anderson (1957) mapped the Baker quadrangle which covers the southern half of 3 and the North Fork quadrangle which includes the northern part of 3.

The subsection is drained by the Lemhi River to the south and the Salmon River to the North. The Beaverhead (south) and Bitterroot Mountains (north) make up most of the region. Elevations exceed 10,000 feet in the subsection. Several periods of Alpine glaciation have affected the area.

The major ridge former and mountain builder in the region is a thick sequence of Precambrian quartzites. Float from batholith outliers (quartz monzonite, Ki3c) was observed by Anderson in the Baker quad and one fair sized outcrop of batholith was mapped in the North Fork quad. Several Tertiary dikes (rhyolites mostly) cut the Precambrian rocks.

The importance of the quartzites as ridge formers is discussed by Anderson (1957) as follows.

Belt Series (Precambrian)

Distribution

"The metamorphosed Beltian rocks are the main mountain builders (Pls. 2 and 3) and compose much of the Beaverhead Range within the quadrangle as well as for some miles to the northwest and southeast. They also form the upper part of K Mountain (Pl. 5, A), the outlier of the Lemhi Range which lies within the southwest corner of the guadrangle. The series passes beneath the younger rocks at the base of the ranges and probably forms the floor upon which rest most, if not all, of the Tertiary formations. It is more widespread than any of the local formations."

Character and Composition

"The Belt rocks show some variation in lithologic character, but the variation is not so much a reflection of changes in rock composition as in differences in degree

> of metamorphism. The rocks along the front of the Beaverhead Range have been more highly deformed and more intensely metamorphosed than those on the higher ridges near the Continental Divide and the country to the east. Their appearance is, therefore, materially different from the less highly metamorphosed rocks which otherwise characterize the series.
>
> Along the front of the Beaverhead Range the Belt series is composed of fine-grained, moderately dark gray, impure quartzitic rocks, usually showing some foliation and in places transition into phyllite. Bedding is generally obscured, but most of the beds appear to be a few inches thick, and in some places where the rock is more highly siliceous the beds are much thicker and the rock more massive with less evidence of foliation. In general, the foliation is more pronounced than the bedding and in places obscures the bedding completely. From a distance the foliation may give the appearance of bedding and unless closely checked may be interpreted incorrectly. The cleavage of these foliated rocks has a marked influence on the topography and tends to control slopes and outcrops (Pl. 3, B). It does not, however, prevent the quartzitic beds from forming bold, rough outcrops in canyon walls. The cleavage is more pronounced in the rocks near the base and lower flank of the range than toward and along the crest and is more evident in rocks near the north end of the quadrangle than farther south. With decreasing foliation toward the crest of the range and toward the southeast, the rocks become lighter-colored and the bedding more distinct, with beds in places appearing alternately pale green and light gray, more siliceous beds alternating with more argillaceous beds, more accurately classed as argillites. In places near the Montana line the less metamorphosed beds resemble closely the Belt rocks in the northern part of the state. Toward the southeast the rocks appear more quartzitic as a whole and in part more massive. Across the Lemhi Valley, K Mountain is composed of rather impure but quite massive quartzite." (Anderson, 1957 p. 14)

The quartzites were mapped as undifferentiated in the Baker quad (mpCsud) but subdivided into the Swauger and Lemhi quartzites (mpCs1 and mpCs2) in the North Fork area (for descriptions see Section III, Subsection 7).

The quartzites are structurally complex. Folding is more severe at the base of the Beaverhead Range than at its crest and deformation increases northwards along the base. Tight, overturned folds mark the base of the range but these become more open and gentle, in the higher elevations. A major fault system extends along the crest of the Beaverhead Range. The extent of this fault to the southeast is not known but it extends for many miles to the northwest. A set of

transverse faults have a east-southeasterly trend. Part of these may be compliments of the longitudinal northwest faults but others are probably independent.

Anderson noted several sets of joints in the quartzites, the major ones striking east-southeast, east northeast, and north northwest. The jointing is so perfect is some areas as to produce sheeted structures which look like bedding from a distance.

Subsection 4 - Salmon Lowlands Subsection

The Salmon-Lowlands Subsection contains the Salmon Basin and the terrace covered foothills of the Beaverhead Range. Anderson mapped the Salmon quad (1956). The Salmon Basin is covered by Tertiary Challis Volcanics and Quaternary surficial deposits in marked contrast to the Salmon and Beaverhead Ranges which are formed from Precambrian quartzites.

Several outcrops of undifferentiated Challis volcanoclastic units, the Kirtly Formation (Tmud), the Geertston Formation (Tmmd), and the Kerney Formation (Tmid).

Two ages of glacial deposits are described by Anderson which he called Pre-Wisconsin and Wisconsin. The older deposits are found as terrace gravels capping ridges below 7,000 feet while the Wisconsin deposits are found only above 8,000 feet (Wisconsin glaciation was less intense than the earlier glacial period and only affected the higher elevation). Three sets of reworked glacial material in the form of fluvial terraces fill the Salmon Basin. Williams Lake in the southwest corner of the quad was dammed by a large landslide and several other smaller slides are present in the region.

Structurally the Basin is related to folding in the Salmon Mountains and the Lemhi Range. A large anticlinal fold (the Salmon River Mountain anticline) lies on strike with the eastern edge of the Salmon River Mountains while a large syncline forms part of the lower Lemhi Range (Lemhi Syncline). Between these two large structures, Anderson observed two imperfect major folds (syncline and anticline) which are only partially revealed along the Salmon River and in the foothills of the Lemhis. The role of faulting within the Basin is difficult to access because of the surficial deposits which cover most fault traces.

The actual formation of the Salmon Basin was discussed by Anderson as follows.

Basin Structure

Salmon Basin

"The Salmon Basin is the largest and most obvious structural feature in the area. This basin in which the Carmen formation has accumlated has every aspect of a basin produced by downwarping. Not only do the Carmen

beds dip inward toward the center, but the older rocks around the rim of the basin also dip toward and then pass under the basin. This relationship of the older rocks is particularly noticeable in the case of the Challis volcanics. Even the ridges forming the eastern slope of the Salmon River Mountains slope downward in such a way as to suggest that they too are involved in the downwarp and that the mountain slope passes under and forms the under floor of the basin.

Whatever part faulting may have had in the development of the basin has been effaced by the fill of sedimentary materials. There appear to be no encircling faults at the margin of the basin and thus no evidence that subsidence might have accompanied faulting. It is anticipated, however, that earlier zones of structural weakness may have had something to do with controlling the location and shape of the basin.

The total subsidence involved in the downwarping is probably well over a mile. The fill, once thicker than now, probably does not exceed 2,000 feet and may be somewhat less.

The origin of the basin must be ascribed to crustal warping."

Age of Basin

"As the broad anticline in the Challis volcanics is involved in the crustal warping, the subsidence must have started shortly after the post-Challis folding, faulting, and igneous intrusion and continued through the lower Miocene, in part contemporaneous with the accumulation of the Carmen formation. Dips of the beds around the margins of the basin indicate that some subsidence continued after the accumulation of the basin materials, probably in response to more recent crustal movements. The basin may be dated as mainly lower Miocene." (Anderson, 1956 p. 51)

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APPENDIX IA

STRATIGRAPHIC SECTION AND LEGEND

FOR THE GEOLOGIC MAP FOLIO

WHICH IS INCLUDED IN ORDER NO. 1032-R4-74

ERA	SYSTEM	SERIES		STRATIFIED ROCKS	MAP SYMBOL	
	Quaternary	Recent		Palouse Fm.	Qw	
		Pleistocene	upper	Caldwell/Nampa Sed.	Qpus	Undiff. Qps
			middle	Ten mile Gravel	Qpls	
				Bruneau Fm.	Qpme	
				Bruneau Fan Gravels	Qpms	
			lower	Idaho Fm.	Tpd	
			general	Undiff. glacial dep.	Qg	
				Undiff. deposits other than glacial	Qa	
		Pliocene	upper	Donkey Fanglomerate	Tpud	
Cenozoic			middle	none		
			lower	none		
			general	none		
		Miocene	upper	Kirtly Fm.	Tmud	
			middle	Geerston Fm.	Tmmd	
			lower	Kerney Fm.	Tmld	
			general	none		
	Tertiary	Oligocene	general	none		
		Eocene	general	none		
		Paleocene	general	Beaverhead Fm.	Tpag	
		Tertiary	general	Payette Fm.	Tmd	
	Cretaeceous	general		none		
Mesozoic	Jurassic	general		none		
	Triassic	general		none		
	Mesozoic	general		none		

	Permian	general	none	
	Pennsylvanian	upper	none	
		middle	Wood River Fm.	Pm
		lower	Copper Basin Sequence	PL
		general	White Knob ls.	Mu
	Mississippian	general	none	
	Devonian	upper	Three Forks ls. Grand View dolomite	Du1 Du2
		middle	Jefferson dolomite	Dm
		lower	none	
		general	none	
Paleozoic	Silurian	upper	none	
		middle	Laketown dolomite	Sm
		lower	none	
		general	Trail Creek Fm.	S
	Ordovician	upper	Saturday Mt. Fm.	S
		middle	Kinnikinnic qtz.	Om
		lower	Ramshorn slate	OL
		general	Phi Kappa Fm.	O
	Cambrian	upper	none	
		middle	none	
		lower	Bayhorse dolomite ?	Cl
		general	none	
	Precambrian	general	Missoula Group=Libby Fm.?Stripped Peak Fm.	pC14
			Wallace Fm.	pC13
			Ravalli Group=St. Regis Fm./Revett qtz./Burke Fm	pC12
			Prichard Fm.	pC11

ERA	SYSTEM	SERIES	METAMORPHIC ROCKS	MAP SYMBOL
Cenozoic	Quaternary	general	none	
	Tertiary	general	none	
	Cretaeceous	general	none	
Mesozoic	Jurassic	general	none	
	Triassic	general	Lucile slate	mTrs
			Martin Bridge ls.	mTrc
	Permian	general	none	
	Pennsylvanian	general	none	
	Mississippian	general	none	
Paleozoic	Devonian	general	none	
	Silurian	general	none	
	Ordovician	general	none	
	Cambrian	general	Garden Creek phyllite	C
Precambrian	Precambrian-general		Missoula Group	mpC14
			Wallace Fm.	mpC13
			Ravalli Group	mpC12
			Prichard Fm.	mpC11
			Yellowjacket Fm.	mpCY
			Hoodoo qtz.	mpCHo
			East Fork Fm. mpCsE, Hyndman Peak Fm. mpCHy > Undiff.	mpCsud
			Swauger qtz. mpCs1, Lemhi qtz. mpCs2 > Undiff.	mpCsud
			Apple Creek phyllite	mpCs3
			Boyle Ranch Sequence	mpCs
			Undiff. Precambrian gneisses and schists.	mpC1
			Augen gneiss	m36

Riggins Group	Squaw Creek schist	$m_2 4$
	Berg Creek Amphibolite	$m_2 3$
	Lightening Creek schist	$m_2 2$
	Fiddle Creek schist	$m_2 1$
Other Riggins metamorphics	Amphibolite	$m_3 5$
	Gneiss complex	$m_3 4$
	Qtz-albite gneiss	$m_3 3$
	Marble	$m_3 2$
	Quartzite	$m_3 1$

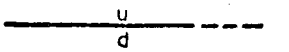
ERA	SYSTEM	SERIES	INTRUSIVE IGNEOUS ROCKS		MAP SYMBOL
Cenozoic	Quaternary	Pleistocene	general	none	
	Tertiary	general	Dike/sill porphyry intrusives	Dacite/ rhyolite	Ti3
				Andesite	Ti2
				Basalt	Ti1
				Undiff.	Ti
			Intrusive stocks and batholiths		Tei3
Mesozoic	Cretaeceous	general	Idaho batholith	Qtz. diorite	Ki2a
				Granodiorite	Ki3b
				Qtz. monzonite	Ki3c
				Leucocratic qtz. monzonite	Ki3a
				Undiff. Idaho batholith	Ki3
				Trondhjemite	Ki3t
			Border zone gneisses and ortho-gneisses within the batholith area	Foliated qtz. monzonite	Ki3cf
				Migmatite	mKi3M
				Qtz. diorite gneiss	mKi2a
				Undiff. border zone	mKi3
				Ortho-gneisses	
			Rock types in the batholith which are only used for the road log of Larsen and Schmidt, 1958.	Granodiorite, Cascade type	Ki3bd
				Granodiorite, Atlanta type	Ki3be
				Fine- textured qtz. monzonite	Ki3cF
				Fine dark qtz. monzonite	Ki3cg
				Coarsw qtz. monzonite	Ki3ch
	Jurassic	general	Metamorphosed intrusives in the Seven Devil,s volcanics		Ji2a
	Triassic	general	none		

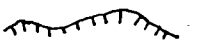
ERA	SYSTEM	SERIES	INTRUSIVE IGNEOUS ROCKS	MAP SYMBOL
	Permian	general	none	
	Pennsylvanian	general	none	
	Mississippian	general	none	
Paleozoic	Devonian	general	none	
	Silurian	general	none	
	Ordovician	general	none	
	Cambrian	general	none	
Precambrian	Precambrian-general		Anorthosite	mpCan
			Undiff. Precambrian metamorphosed intrusives	mpCil

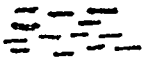
ERA	SYSTEM	SERIES	EXTRUSIVE IGNEOUS ROCKS	MAP SYMBOL
		Recent	Recent lava flows	Qre
	Quaternary	Pleistocene	none	
		general	Undiff. Snake River basalts	Qpue
Cenozoic				
		Pliocene- general	Banbury basalt	Tpme
			Idavada volcanics	Tple
		Miocene- general	Challis volcanics: Yankee Fork rhyolite	Tme3
	Tertiary	Oligocene-general	Challis volcanics: Germer tuffaceous member	Tov
		Eocene- general	Challis volcanics: Latite-andesite member	Tee2
		Paleocene- general	none	
		Tertiary- general	Undiff. Challis volcanics	Tev
			Casto volcanics(altered Challis volcanics)	Pv
			Basalts in the Challis volcanics	Tel
			Kamiah volcanics	Toel
	Cretaeceous	general	none	
Mesozoic	Jurassic	general	none	
	Triassic	general	Seven Devil's volcanics (metamorphosed to greenstones)	mTrv
Paleozoic		general	There are no Paleozoic extrusive rocks	
Precambrian		general	There are no Precambrian extrusive rocks	

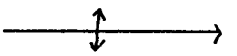
MAP SYMBOLS

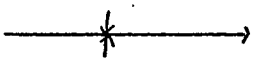
Formation or Unit contact (dashed where approximately located)

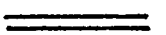
Normal or Reverse Fault (dashed where approximately located) 

Thrust Fault (teeth on upper plate) 

Shear zone 

Anticline or antiform 

Syncline or synform 

Zeolite/Greenschist Isograde 

ADDENDA

Appendix 1B:

Day, N., Hall, W.B., and John Bond, 1974, Application of remote sensing and orbital imagery to resource evaluation and land use decision making: STAR project final report, State of Idaho. Project 41-076

Linear Map compilations:

Linear Maps: Scale 1/250,000 (AMS), compiled by Norman Day, currently with the U.S. Forest Service, Orofino,Idaho.

The Geology of Section IV:

Day, N., 1973, Tectonic map of Idaho from ERTS imagery (Abs): G.S.A. Abstracts with programs, V.5,No.7, p.594.

APPENDIX IB

EXPLANATION OF LINEAR MAP OF IDAHO FROM ERTS IMAGES

LINEAR MAP OF IDAHO FROM ERTS IMAGES

Introduction

This is a progress report for the S.T.A.R. research project "Application of Remote Sensing and Orbital Imagery to Resource Evaluation and Land-Use Decision Making". As stated in the original proposal, the ERTS image research is state-wide in scale and concentrates on three topics:

1. Delineation of imagery linears and structural belts to determine to what degree they are related to, or diagnostic of, mineralized areas and geothermal energy sites.

2. Identification of major rock types and unconsolidated materials, in particular as they might relate to appropriate land-use considerations such as mineral resources, aggregate sources, recreational areas, and primitive areas.

3. Evaluation of multispectral imagery and repetitive coverage to improve interpretations through monitoring seasonal changes and sun-angle effects.

The remainder of this report will discuss the level of progress made in each of the above topics.

DELINEATION OF IMAGERY LINEARS AND STRUCTURAL BELTS

This section was initiated first and is the nearest to completion. To date, a preliminary linears base map of Idaho (1:250,000) has been constructed and shows all major rectilinears and curvilinears which can be recognized from band 5 ERTS imagery. From the original state-wide linears base map a positive correlation between mineral and geothermal localities with linears has been demonstrated and a linear domain map has been produced.

Linears Base Map

The major part of this project was carried out on band 5 paper prints in 15" X 15" format. More recently selected images from bands 4, 6, and 7 have been applied to the study. Images on hand for this study were all taken in September and October of 1972; ordered repetitive coverage has not been received at this time.

Each image was carefully examined (with North up) for the purpose of locating all major linear elements of over one mile in length. Once located on the image the linears were transferred by visual inspection to Army Map Service topographic base maps of a scale 1:250,000.

In general, the linears observed in this study range from the one mile minimum to nearly 100 miles and in density from as few as four to more than 60 per 500 square miles (Plate 1 -- Linear Map).

A second examination of the ERTS images is currently underway to compensate for the poor quality of some of the original images, the need to use different image orientations when searching for linears and the capability for certain MSS bands to show more reliable structural information than other bands. This should be completed by the termination date for this grant.

Linear Domain Map

In an effort to organize the general random nature of the statewide linears, all linears within the state have been carefully analyzed and grouped according to linear patterns which include length, shape, direction, and density. This process reveals at least ten separate linear domains within Idaho. (Plate 2 -- Linear Domain Map) An attempt was then made to determine what factors produced the observed similarity of linear patterns. A summary of each domain follows and will conform to the format given below.

#____ Name of Domain

Primary linear trends and lengths.
Secondary linear trends and lengths.
Density of linears (Extreme, high, moderate, low, and sparse)
General bedrock found in the domain.
Remarks

1 Snake Plain Downwarp

Primary trends -- NNW & NW from 10 to 15 miles in length.
Secondary trends -- NNE & NW treater than 5 miles.
Density -- High.
Bedrock -- Tertiary and Quaternary volcanics .
Remarks -- NNW & NW linears appear to represent tensional features which parallel the basinal axis.

2 Idaho-Utah-Wyoming overthrust belt

Primary trends -- NNW & NW from 10 to 15 miles in length.
Secondary trends -- NW 5 miles or less.
Density -- Moderate.
Bedrock -- Paleozoic and Mesozoic sediments.
Remarks -- The NNW & NW linears reflect thrusting; some of the NE linears demonstrate tear faults.

3 Snake Plain

Primary trends -- NNW & NE from 15 to 20 miles in length.
Secondary trends -- None.
Density -- Sparse.
Bedrock -- Quaternary volcanics.
Remarks -- Linears may represent rift zones for fissure eruptions.

4 Yellowstone Plateau

Remarks -- Due to poor image quality, small area and band 5 coverage only, structural data is adequate only to show that this domain is different from those surrounding it.

5 Idaho-Montana fold belt

Primary trends -- No single directional trend predominates.
Secondary trends -- Not applicable.
Density -- Moderate
Bedrock -- Paleozoic and Mesozoic sediments.
Remarks -- Some of the NE linears appear to represent tear faults.

6 Sawtooth-Challis region

Primary trends -- NW & NE 5 miles or less in length.
Secondary trends -- NW & NE greater than 10 miles.
Density -- Extreme.
Bedrock -- Paleozoic and Mesozoic sediments, Cretaceous granite and Tertiary volcanics.
Remarks -- The short NW & NE trending linears are interpreted as representing two orthogonal joint sets.

7 The Weiser Embayment

Remarks -- Some possible plunging structures are noted, although linears are nearly absent in the Tertiary basalt.

8a Idaho Batholith (South)

Primary trend -- NNE from 5 to over 35 miles in length.
Secondary trends -- NW from 5 to 10 miles.
Density -- Moderate.
Bedrock -- Cretaceous granite and related rocks.
Remarks -- Some of the NNE linears may be significant strike slip faults.

8b Idaho Batholith (North)

Primary trend -- NE from 10 to over 40 miles in length.
Secondary trend -- NW 5 to 20 miles.
Density -- High.
Bedrock -- Cretaceous granite and related rocks.
Remarks -- Curvilinears suggest drag features associated with the trans-Idaho discontinuity.

9 Clearwater Embayment

Primary trend -- NW from 5 to 15 miles
Secondary trend - NE with linears greater than 20 miles in length.
Density -- Low to moderate.
Bedrock -- Tertiary basalt.

10 Idaho-Montana structural belt

Primary trend - WNW with linears greater than 50 miles in length.
Secondary trends -- NE & NW from 5 to 10 miles.
Density -- Moderate.
Bedrock -- Precambrian sediments and Cretaceous granite.
Remarks -- The NE & NW trending linears are more numberous, but apparently secondary to the WNW linears. WNW linears represent large strike slip faults; some lateral movement is also suggested by the NE & NW trending linears.

These domains show a varying coincidence with bedrock type, bedrock age, and structural province. Plate 3 shows the relationship between linear domains and observed structural trends in Idaho. Coincidence of domains with bedrock type and bedrock age will be discussed under the section dealing with bedrock identification from ERTS.

Mineral and Geothermal Relationships with Linears

Another aspect of this investigation has been to determine what relationship, if any, exists between mineral and geothermal localities with linear trends.

A statewide compilation of all known mineral localities onto A.M.S. maps is being supported by private industry and nearing completion. To date, the mineral inventory includes all of Idaho above 44° North latitude. Therefore, the following mineral discussion and tables are only pertinent to the area north of the Snake Plain. Comparison of mineral localities with ERTS linears demonstrates that 77% of known mineral sites are on mapped linears. Mineral localities occur primarily along NW and NE trending linears. Secondary trends for mineral occurrence are NNE and NNW. Please

note that there are no data yet confirming that there are preferred mineralizing trends. The higher persistence of minerals with certain linear trends may only reflect the higher density of linears falling in these trends. A statistical study is in progress to determine this association and if preferred trends exist.

The S.T.A.R. study demonstrates a strong correlation between both geothermal sites and mineralized localities with linears. Primary results of this portion of the study are shown by two tables. Table 1 is an evaluation by A.M.S. sheet of the association of linear trends with mineral sites (north of Snake Plain) and geothermal sites (statewide). Table 2 presents by linear domains the association of linear trends with mineral sites (north of the Snake Plain) and geothermal localities (statewide). Data has been normalized (to 100%) for tabulation.

At this point it should be mentioned that just because some mineral and geothermal sites do not fall along observed linears it should not be assumed that they do not occur along some kind of undetected fracture. These linears may be undetected because they are either very old fractures and have had no recent movement or amplification by differential erosion or they are very recent features which have not been excavated by erosional processes.

Comparison of all known geothermal springs and wells in Idaho with ERTS linears show that 80% of all reported geothermal sites occur along linears. Geothermal sites are found primarily along NE, NW, and NNE trending linears.

The degree of correlation between mineral and geothermal sites with linear trends appears controlled in part by bedrock. When considering mineral localities, the greatest degree of coincidence is found primarily in areas of Precambrian metasediments; the poorest coincidence, and least occurrence, is found in areas where volcanic rock is present (Figure 1). The highest degree of correlation for geothermal sites is in areas of intrusive crystalline rocks. The poorest correlation, again, is in areas of volcanic and sedimentary rocks (Figure 2).

IDENTIFICATION OF MAJOR ROCK TYPES

This second portion of the study, dealing with identification of major rock types and unconsolidated materials, has received lesser attention to date. Preliminary attempts at bedrock differentiation indicate that only geologic mapping on the grossest scale is possible from ERTS images on hand. Listed below are several comments concerning bedrock recognition from ERTS imagery. Further studies will be completed when compositing of images of different bands is possible.

1. Three factors have limited our scope and now give us the opinion that ERTS is not a preferred medium for bedrock mapping in Idaho; these

are: a). The major portion of this study to date was carried out on band 5 images which emphasize vegetation cover, not rock. b). Most of Idaho above 43° 30N latitude is thickly vegetated. c). Large portions of Idaho are underlain by rock types (intrusive and extrusive igneous) which appear homogeneous on aerial photography.

2. There is some indication that linear patterns can delineate areas of similar bedrock, but only on a very large scale as is indicated by the association of linear domains with bedrock type (Plage 4).

3. Valley alluvium generally poses no problems in its identification from ERTS imagery as long as there is sufficient topographic difference between it and the surrounding bedrock.

4. As yet, no one kind of bedrock had been observed which has a diagnostic, four band, spectral response that will consistantly differentiate it from other rock types throughout Idaho. However, it is believed that by using a densitometer and/or a color compositor and concentrating on rock types in southern Idaho it may be possible to "fingerprint" certain rock types.

EVALUATION OF MULTISPECTRAL IMAGERY AND REPETITIVE COVERAGE

Although all four bands of ERTS imagery for selected areas were examined, it was evident early in the project that the several bands provided differing amounts and types of structural and geologic information. Impressions of the four multispectral bands used for this study, as they relate to Idaho, follow:

Band 4: 0.5 to 0.6 micrometers (green wavelength) - This band provides the poorest structural information of all MSS bands. However, it excells in delineating sand dunes or snow cover, it may discriminate the depth and/or turbidity of standing bodies of water. It appears good for making cultural and vegetation studies. Its overall appearance is "muddy" and low contrast.

Band 5: 0.6 to 0.7 micrometers (lower red wavelength) - Band 5 provides good to very good overall geologic and structural information in areas of little vegetation and few cultural features and is the best for displaying topographic and cultural features.

Bands 6 and 7: 0.7 to 0.8 and 0.8 to 1.1 micrometers (near infrared wavelengths - These bands prove excellent for determining structural lineaments and providing geologic information in densely populated and sensely vegetated areas. In semi-arid and arid regions, bands 5, 6, and 7 were very nearly equivalent in their ability to detect structural lineaments. It appears that band 6 is the best all-around spectral band for most geologic studies in Idaho; it penetrates and de-emphasizes thick vegetal cover or dense cultural features and provides excellent water course delineation without the photographic "bleeding" and consequent shape distortion so common band 7.

To demonstrate the ability for bands 6 and 7 to detect linears that bands 4 and 5 cannot, an area in the densely forested mountains of north-central Idaho was chosen as a test area (Plate 2). First, bands 4 and 5 were examined and all detectable linears were plotted onto a base map. Next, bands 6 and 7 were examined in the same manner as bands 4 and 5. From this process, two areas where linears were not detected on bands 4 and 5 were located (Plate 2). These areas are both characterized by low relief, dense vegetation, and thick soils. Linears in the southern are reflect a series of low, sub-parallel ridges which may represent dikes. The northern zone of curvilinears may reflect drag features indicating left-lateral movement along a shear zone which corresponds to Yates' Trans-Idaho Discontinuity; a more detailed study of this topic is currently underway and will be presented at the Geological Society of America's Cordilleran sectional meeting in March, 1974. Several very long NE trending linears were observed and on underflight photography are shown to be zones of short sub-parallel linears. Due to the lack of repetitive coverage, the portion of this original program dealing with the effects of seasonal changes and sun angle on geologic interpretations could not be implemented by the study's midpoint.

CONCLUSIONS AND SUGGESTIONS

1. A massive compilation of large scale structural trends is both possible and practical through interpretation of the multi-spectral images transmitted from the ERTS-1 satellite.

2. A large proportion of all determinable linears seem to be shown on ERTS images so that meaningful maps delineating areas with significantly different linear densities and trend distributions can be made.

3. Experience has shown that it is possible to make a more complete and reliable linears map and structural interpretation from high-altitude, stereo, vertical airphotos than from ERTS imagery. However, the cost for such photography is much greater because of the large number of prints needed to cover the same area. Also the time needed to evaluate all of the underflight photos would be orders of magnitude greater than for ERTS evaluation.

4. Almost no satisfactory geologic mapping except on the broadest scale has been possible from ERTS images (primarily band 5) alone. In this study, the satellite scale nonstereo images are not at all competitive with normal stereo vertical airphotos.

5. After examining all four MSS bands, it appears that band 6 presents the most structural information with the least cultural and vegetal interference.

6. This study is demonstrating a strong correlation between both mineral localities and geothermal sites with linear trends.

7. To date, this project has produced information leading to three formal talks given at professional geologic meetings with subsequent publication of two abstracts; the two abstracts are found attached.

8. Prior to final production of the state linears map all linears should be statistically analyzed (length, azimuth, and density) to determine major structural trends and verify the linear domains. Their statistical treatment is also needed to establish if there are preferred trends for mineralization and geothermal activity.

TABLE I -- (BY A.M.S. SHEETS) *

Linear Trends

A.M.S. Sheet	Sites	WNW	NW	NNW	N	NNE	NE	ENE	EW	Off	% Total
Baker Geothermal	(14)	7	-	7	-	14	21	-	-	50	100
Baker Minerals	(51)	-	31	4	-	2	4	-	-	59	100
Boise Geothermal	(40)	-	8	-	-	13	30	-	-	50	100
Boise Minerals		Survey not complete									
Challis Geothermal	(45)	-	-	7	-	51	20	11	7	4	100
Challis Minerals	(270)	6	26	11	9	19	12	.3	6	10	100
Dillon Geothermal	(1)	-	-	-	-	-	100	-	-	0	100
Dillon Minerals	(107)	-	7	4	1	10	32	2	-	44	100
Driggs Geothermal	(11)	-	-	-	-	9	64	-	-	27	100
Driggs Minerals		Survey not complete									
Dubois Geothermal	(8)	-	13	-	-	25	38	-	-	25	100
Dubois Minerals	(140)	16	19	6	2	17	11	1	1	26	100
Elk City Geothermal	(10)	20	-	-	-	-	40	-	-	40	100
Elk City Minerals	(312)	3	21	14	3	12	26	2	3	20	100
Grangeville Geothermal	(5)	-	-	40	-	20	20	-	-	20	100
Grangeville Minerals	(84)	1	-	40	14	-	11	-	-	33	100
Hailey Geothermal	(60)	-	7	17	-	10	45	10	7	5	100
Hailey Minerals		Survey not complete									
Hamilton Geothermal	(9)	-	22	-	-	11	44	22	-	0	100
Hamilton Minerals	(37)	5	16	-	3	8	35	-	-	32	100

* Indicate maps in this report.

A.M.S. Sheet	Sites	WNW	NW	NNW	N	NNE	NE	ENE	EW	OFF	% TOTAL
	Geothermal(4)	-	-	-	50	50	-	-	-	0	100
Idaho Falls											
	Minerals Survey not complete										
	Geothermal(19)	-	21	-	-	-	58	-	-	21	100
Jordan Valley											
	Mineral Survey not complete										
	Geothermal(22)	5	45	-	-	5	27	-	-	18	100
Pocatello											
	Minerals Survey not complete										
	Geothermal(13)	-	15	23	-	31	15	8	-	8	100
Preston .											
	Minerals Survey not complete										
	Geothermal(0)	No reported localities									0
Pullman											
	Minerals (37)	-	5	27	3	19	14	-	-	32	100
	Geothermal(0)	No reported localities									0
Sandpoint											
	Minerals (159)	-	35	9	1	21	14	-	-	20	100
	Geothermal(0)	No reported localities									0
Spokane											
	Minerals (73)	7	30	21	-	15	12	-	3	12	100
	Geothermal(58)	-	48	5	7	2	17	-	-	21	100
Twin Falls											
	Minerals (0) Survey not completed										
	Geothermal(0)	No reported localites									0
*Wallace											
	Minerals (89)	7	24	1	1	7	43	-	1	17	100

* Indicate maps in this report.

*Coeur d'Alene district absent from these figures.
-Minerals surveyed include nonplacer gold, silver, copper, lead, zinc, iron moly and tungsten.

Table 1. Association of mineral and geothermal sites with linears. Figures indicate the % of mineral localities and geothermal sites which fall along linears within the given directional catagories for individual A.M. S. sheet.

TABLE II -- (BY LINEAR DOMAINS)

Linear Trends

Linear Domains	Sites	WNW	NW	NNW	N	NNE	NE	ENE	EW	OFF	% TOTAL
#1 Geothermal	(117)	_	30*	3*	3	5+	28+	-	-	31	100
#1 Mineral		Survey not complete									
#2 Geothermal	(46)	2	26*	7*	-	13+	33+	2	-	17	100
#2 Mineral		Survey not complete									
#3 Geothermal	(4)	-	-*	-	50	50	-	-	-	0	100
#3 Mineral		Survey not complete									
#4 Geothermal		Data for area not yet determined									
#4 Mineral											
#5 Geothermal	(9)	-	11	-	-	33	33	-	-	22	100
#5 Mineral	(247)	9	14	5	2	14	20	2	4	34	100
#6 Geothermal	(105)	-	4*+	12	-	28	34*+	10	7	5	100
#6 Mineral	(270)	6	26*+	11	9	19	12*+	.3	6	10	100
#7 Geothermal	(14)	7	-	7	-	14	21	-	-	50	100
#7 Mineral	(51)	-	31	4	-	2	4	-	-	59	100
#8 a & b Geothermal	(24)	8	8+	8	-	8*	38*	8	-	21	100
#8 a & b Mineral	(433)	3	16+	18	5	9*	24*	1	.2	23	100
#9 Geothermal		No reported localities									
#9 Mineral	(37)	-	5*	27	3	19	14+	-	-	32	100
#10 Geothermal		No reported localities									
#10 Mineral	(321)	3*	31+	9	1	16	22+	-	1	17	100

* Indicates primary linear trends within each domain.
\+ Indicates secondary linear trends within each domain.

Figure 1 -- Mineral--Bedrock Associations

	A.M.S. Map	% Coincidence	Dominant Bedrock of A.M.S. Map Area
	Challis	90	K bath. & T volc.
Good Correspondence	Spokane	88	Pe sediments
	Wallace	88	Pe sediments
	Pullman	68	T volc.
	Grangeville	67	T volc. & K bath.
Poor Correspondence	Dillon	56	T volc. & Pe seds.
	Baker	41	T volcanics

Figure 2 -- Geothermal-Bedrock Associations

	A.M.S. Map	% Coincidence	Dominant Bedrock of A.M.S. Map Area
	Hamilton	100	K bath. & Pe seds.
Good Correspondence	Challis	96	K bath. & T volc.
	Hailey	95	K bath. & Q volc.
	Dubois	75	Pz sediments
Poor Correspondence	Driggs	73	T volc. & Pz-Mz seds.
	Boise	50	Q volc. & K. bath.

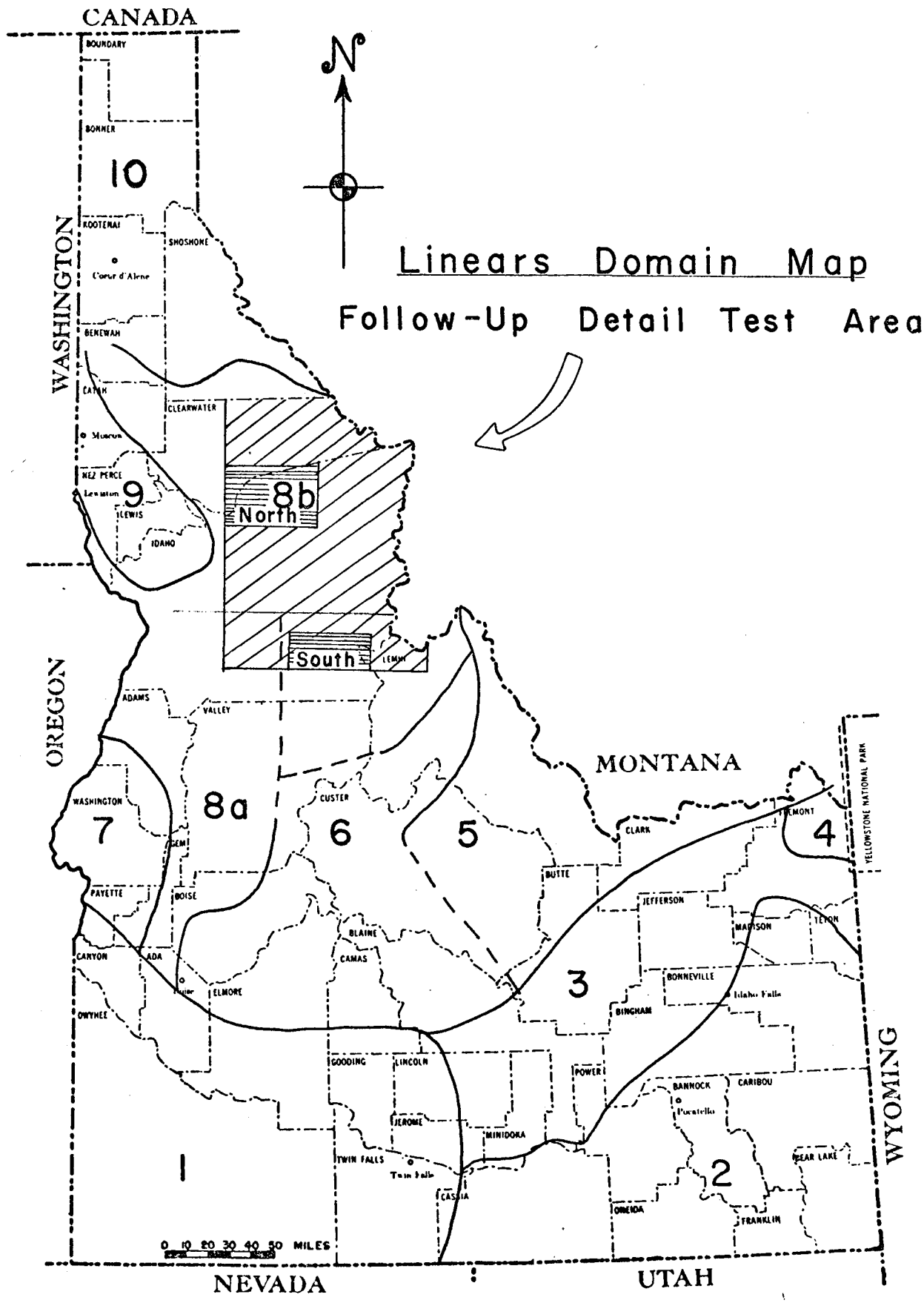

CANADA
BOUNDARY
BONNER
10
KOOTENAI
SHOSHONE
Coeur d'Alene
BENEWAH
WASHINGTON
LATAH
CLEARWATER
Moscow
NEZ PERCE
Lewiston
9
LEWIS
IDAHO
8b
North
South
LEMHI
Linears Domain Map
Follow-Up Detail Test Area
ADAMS
VALLEY
OREGON
MONTANA
WASHINGTON
7
8a
CUSTER
6
5
CLARK
FREMONT
4
YELLOWSTONE NATIONAL PARK
GEM
PAYETTE
BOISE
BUTTE
JEFFERSON
MADISON
TETON
BLAINE
CAMAS
CANYON
ADA
ELMORE
3
BONNEVILLE
Idaho Falls
BINGHAM
OWYHEE
GOODING
LINCOLN
POWER
BANNOCK
CARIBOU
Pocatello
JEROME
MINIDOKA
TWIN FALLS
Twin Falls
1
2
BEAR LAKE
CASSIA
ONEIDA
FRANKLIN
WYOMING
0 10 20 30 40 50 MILES
NEVADA
UTAH

PLATE 2

IB-14

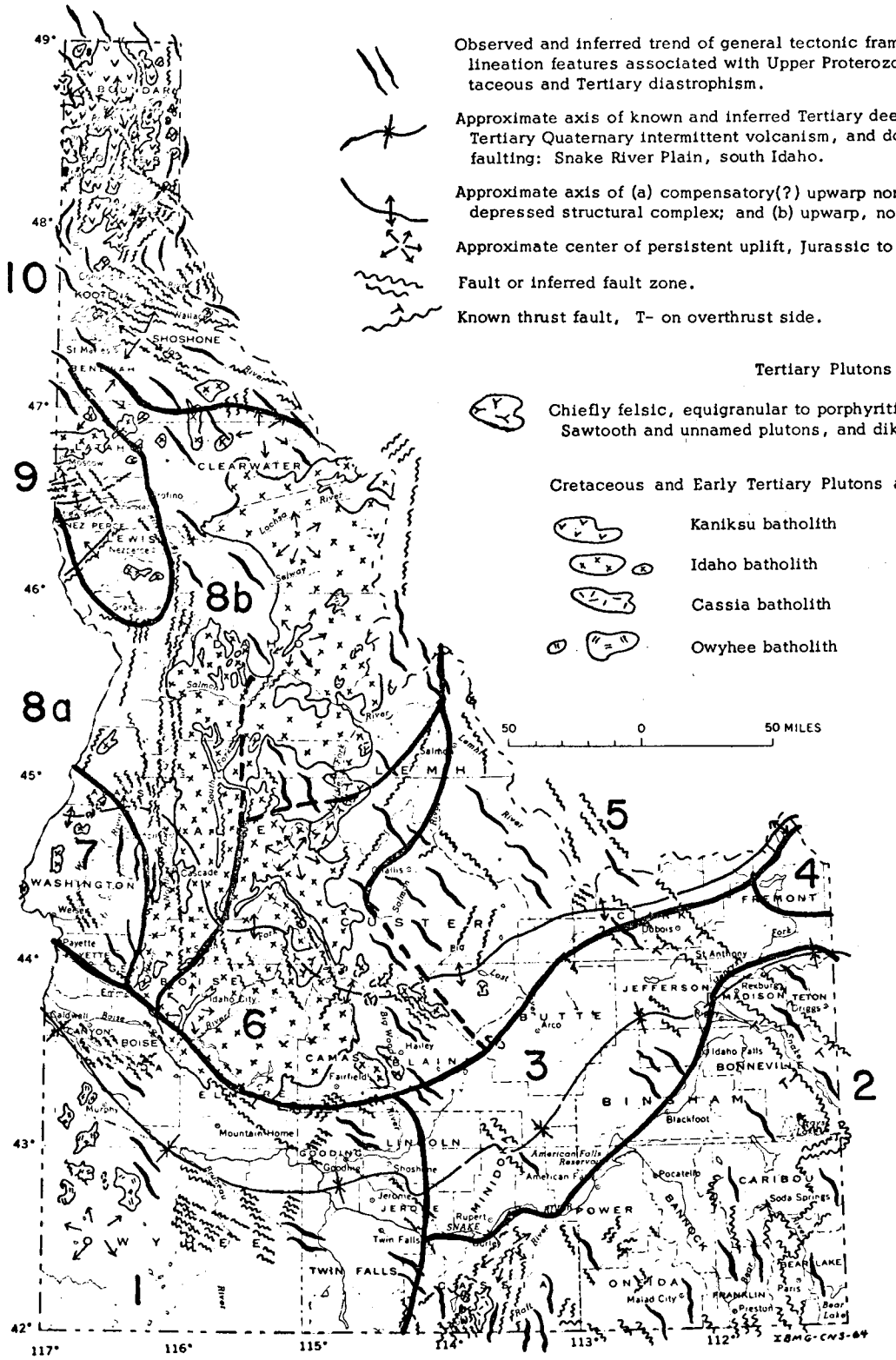


PLATE 3 PRELIMINARY TECTONIC MAP OF IDAHO

by

C. N. SAVAGE

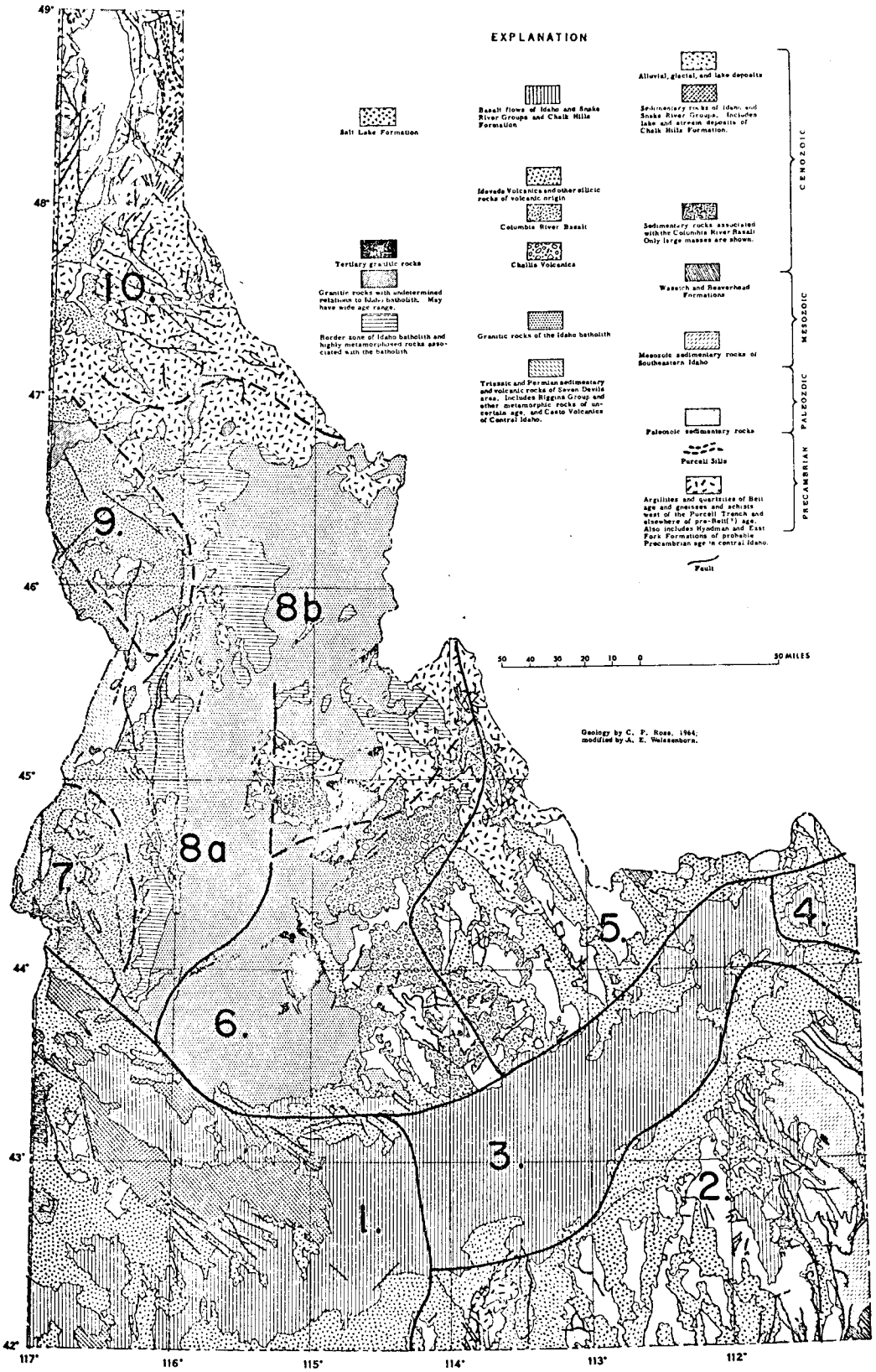
June 1964 IB-15

Linears Domains

by

N.F. Day

December 1973


EXPLANATION
Alluvial, glacial, and lake deposits
Basalt flows of Idaho and Snake River Groups and Chalk Hills Formation
Sedimentary rocks of Idaho and Snake River Groups. Includes lake and stream deposits of Chalk Hills Formation.
Idavada Volcanics and other silicic rocks of volcanic origin
Sedimentary rocks associated with the Columbia River Basalt. Only large masses are shown.
Columbia River Basalt
Chalk Hills Volcanics
Tertiary granitic rocks
Granitic rocks with undetermined relations to Idaho batholith. May have wide age range.
Wasatch and Beaverhead Formations
Granitic rocks of the Idaho batholith
Border zone of Idaho batholith and highly metamorphosed rocks associated with the batholith
Mesozoic sedimentary rocks of Southeastern Idaho
Triassic and Permian sedimentary and volcanic rocks of Seven Devils area. Includes Riggins Group and other metamorphic rocks of uncertain age, and Casto Volcanics of Central Idaho.
Paleozoic sedimentary rocks
Purcell Sills
Belt Lake Formation
Argillites and quartzites of Belt age and gneisses and schists west of the Purcell Trench and elsewhere of pre-Belt(?) age. Also includes Hyndman and East Fork Formations of probable Precambrian age in central Idaho.
Fault
CENOZOIC
MESOZOIC
PALEOZOIC
PRECAMBRIAN
50 40 30 20 10 0 50 MILES
Geology by C. F. Ross, 1964; modified by A. E. Weissenborn.
10.
9.
8b
8a
7
6.
5.
4.
3.
2
1.
49°
48°
47°
46°
45°
44°
43°
42°
117°
116°
115°
114°
113°
112°

Geologic Map of Idaho with Linears Domains Superimposed

PLATE 4 (IB-16)